Discovery of Early Cretaceous rocks in New Caledonia; new geochemical and U-Pb zircon age constraints on the transition from subduction to marginal breakup in the Southwest Pacific.

Dominique Cluzel, C.J. Adams, Sébastien Meffre, H. Campbell, Pierre Maurizot

Discovery of Early Cretaceous rocks in New Caledonia; new geochemical and U-Pb zircon age constraints on the transition from subduction to marginal breakup in the Southwest Pacific.

D. Cluzel[1-2], C.J. Adams[3], S. Meffre[4], H. Campbell[5], and P. Maurizot[6]

[1] Pôle Pluri-disciplinaire de la Matière et de l'Environnement-EA 3325, Université de la Nouvelle-Calédonie, BP R4, 98851 Noumea cedex, New Caledonia. dominique.cluzel@univ-nc.nc

[2] Institut des Sciences de la Terre d'Orléans-UMR 6113, Universities of Orleans and Tours, France

[3] GNS Science, Private Bag 1930, Dunedin, New Zealand.

[4] ARC Centre of Excellence in Ore Deposits, the University of Tasmania, Hobart

[5] GNS Science, PO box 30368, Lower Hutt, New Zealand.

[6] BRGM-SGNC, Noumea New Caledonia

Abstract

New U-Pb dating of detrital zircon and geochemical features of Permian-Mesozoic arc-derived volcanic rocks and volcaniclastic turbidites (greywackes), when compared to the volcanic rocks associated with unconformable Late Cretaceous shallow-water sediments, reveal that subduction in New Caledonia, once thought to be extinct in the Late Jurassic (ca. 150 Ma), was still active at least from ca. 130 Ma to 95 Ma. The accumulation of volcanic arc-derived sediments during the late Early Cretaceous suggests that, as in New Zealand, active margin activity went on for a short time in spite of the assumed subduction jamming by the Hikurangi plateau at ca.100 Ma. Meanwhile, the rift-related magmatic activity that preceded the marginal breakup migrated eastward: from ca. 130 Ma (130-95 Ma) in eastern Australia, 110 Ma (110-82 Ma) in New Zealand, and finally ca. 89 Ma (89-83 Ma) in New Caledonia, and generated large volumes of silicic magma. In contrast, marginal basins opened synchronously at ca. 83 Ma when the stretched continental crust finally broke out. In general, intraplate and volcanic-arc signatures coexisted in Cretaceous syn-rift magmas. Therefore, the

Australian marginal breakup appears to be the final effect of continuous southward unzipping of Gondwana that interfered with the subduction-modified mantle wedge of the Mesozoic active margin. The occurrence of lateral flow of the upper asthenospheric mantle due to the rapidly eastward migrating Australian plate margin possibly prevented the formation of a volcanic arc at the eastern end of the system.

Keywords

New Caledonia, New Zealand, Australia, Gondwana, Mesozoic, active margin, marginal basin, volcaniclastic turbidite, greywacke, geochemistry, U-Pb geochronology, detrital zircon.

Introduction.

Knowledge of the southern Gondwanaland margin is still uncertain and new evidence based upon accurate dating methods is crucial to better constrain its evolution, in particular for the pre-Late Cretaceous period. Compared to the rest of Gondwanaland, this part of the supercontinent underwent break-off very late in its southward separation, initially in the Central Atlantic (Early Jurassic), then the South Atlantic (upper Early Jurassic), then the Mozambique basin (mid-Cretaceous), then at Australia's southeastern margin (Gippsland basin, lower Late Cretaceous) and finally to the Antarctic Ocean (Campanian ?) and Southwest Pacific (Campanian) (Marzoli et al, 1999; Müller et al., 2000; Storey et al, 1999; Bryan et al., 1997; Gaina et al., 1998a & b; Eagles and König, 2008). The close association of break-off features and continental flood basalt or thick volcano-sedimentary successions gave rise to the hypothesis of sequential mantle plume activation (Wilson, 1997; Bryan et al., 1997; Courtillot et al, 1999; Eagles and König, 2008). However, this interpretation does not fit all the features of the Southwest Pacific where a mantle plume, if one existed, probably interfered with subduction (Schellart et al., 2006). During the Late Cretaceous and Palaeocene, the plate boundary separating Australia from Pacific-related plates, which had been situated close to the main Gondwana landmass since the late Palaeozoic, moved quickly 2000 to 3000 km eastward and then remained in an almost unchanged location, in spite of the numerous Tertiary to Recent tectonic events that occurred. Meanwhile, several marginal basins opened along the Australian margin, and in the process created isolated and elongate slices of thinned continental crust that now form continent-sized submarine plateaus or rises such as the Lord Howe Rise, Challenger and Campbell Plateaux, Chatham Rise, Norfolk Ridge and lesser units (Gaina et al, 1998a,b) (Fig. 1).

The formation of these marginal basins gave rise to several conflicting interpretations: i) marginal rifting triggered by a mantle plume (Bryan et al, 1997); ii) subduction of a spreading ridge (Bradshaw, 1989); iii) slab capture (Luyendik, 1995); or iv) marginal basin opening controlled by eastward slab roll-back and arc migration in a continuously converging environment (Lister and Etheridge, 1989; Veevers et al., 1991, Veevers, 2000, 2004; Cluzel et al., 1999; Cluzel et al., 2001; Betts et al., 2002; Schellart et al., 2006; Cluzel et al., 2010). However, a possible connection with Gondwana break-up has been rarely considered (i.e. Tulloch et al., 2009) and deciding whether Australia's marginal rifting is related to circum-Pacific evolution (e.g., a supra-subduction feature); or alternatively, a consequence of super-continent break-up, requires a precise knowledge of the timing and geochemical features of the associated magmatic events.

In this paper, we report U-Pb detrital zircon ages and geochemistry of Mesozoic volcaniclastic sediments, volcanics, and cover rocks of New Caledonia that formed contemporaneously with this series of magmatic events. These new data complement a previous study by Adams et al. (2009), which mainly addressed the pre-Late Cretaceous evolution of the New Caledonia sector of the Gondwana active margin. We focus here on the first evidence for Early Cretaceous rocks in New Caledonia and present a synthesis of the available chronological data on the Late Cretaceous rift-related magmatic activity and marginal basin opening across the Australian margin, which allows a model for the marginal rifting to be proposed.

Pre-Late Cretaceous geology of New Caledonia

In New Caledonia, Permian to mid-Cretaceous rocks occur in three major tectonostratigraphic units: the Teremba, Koh-Central, and Boghen terranes (Fig.1). These dominantly volcanosedimentary terranes are overlain by unconformable Late Cretaceous terrigenous shallow marine sediments. The three terranes that form the central mountains of New Caledonia are as follows:

i) Koh-Central Terrane: a disrupted, Early Permian (Aitchison et al, 1998) ophiolite suite that occurs locally along the centre of the island. It is comprised of gabbro, dolerite, rare plagiogranite, island-arc tholeiites (IAT) and boninite pillow lavas, and undated chert directly overlying the pillow basalts (the Koh Ophiolite of Meffre et al., 1996). The Koh Ophiolite rocks are in turn overlain by a thick deep-water succession of volcano-sedimentary rocks: black shale, volcaniclastic turbidite (classically referred to as greywacke), radiolarian-bearing siltstone and chert (Meffre et al., 1996). The black shales are several hundred metres thick, whilst volcaniclastic sandstones are generally associated with 20-50% argillite and abyssal chert, giving this terrane a distal and deep-water character. The sandstones are exclusively composed of volcanic lithic (andesite and basalt) and mineral clasts (feldspar, gulfed quartz, amphibole, etc.), while plutonic clasts are generally absent. Without exception, the geochemical and isotopic features of the volcaniclastic sandstone relate to subduction-zone magmatism, with no evidence for contamination by sediment or continental crust rocks. This feature is consistent with the eruption of the source rocks through oceanic or intermediate (thinned) crust (see geochemistry section below) (Meffre, 1995; Adams et al., 2009). Middle Triassic (Anisian), and Late Jurassic (Oxfordian-Kimmeridgian) faunas are correlated with those of the New Zealand Murihiku Terrane (Campbell et al., 1985; Meffre, 1995).

ii) Teremba Terrane: a succession of Late Permian to mid-Jurassic proximal volcaniclastic and volcanic rocks (andesite, dacite and rhyolite). The sedimentary rocks are typically medium-grained volcaniclastic turbidite with only minor (<10%) intercalated argillite, some shallow-water volcaniclastic conglomerate, and rare black shale, a few tens metres thick associated with thin quartzose sandstone. This terrane also contains abundant faunas resembling those of the Murihiku Terrane of New Zealand (Grant-Mackie et al., 1977; Paris 1981, Campbell, 1984; Campbell et al., 1985; Ballance & Campbell, 1993). The mineral, chemical and isotopic composition of greywackes closely resembles that of the Central Terrane and similarly contains only very few detrital zircons (Adams et al., 2009). Thus, they are likely derived from the same volcanic or sub-volcanic source.

iii) Boghen Terrane (the ante-Permian of Paris, 1981): an accretionary complex comprising schistose unfossiliferous and broken volcano-sedimentary rocks (pillow basalt, chert, black shale, sandstone, tuffs, turbiditic greywacke, serpentinite and mafic/ultramafic melange), at a notably higher metamorphic grade (lower greenschist to blueschist facies) than the adjacent terranes. Besides widespread retrograde greenschist facies rocks, the occurrence of lawsonite-glaucophane, epidote-glaucophane, and garnet-glaucophane schists allows an intermediate high-pressure low-temperature gradient to be defined. Peak metamorphic conditions indicate subduction of accreted and melanged material at depth equivalent to ca. 12 kbar at ca. 500°C (Cluzel & Meffre, unpub. data). The disposition of metamorphic isograds indicating westward-increasing grade (Guérangé et al., 1977; Paris, 1981), the orientation and top-to-the-east kinematics of exhumation-related stretching lineation, together suggest westward-plunging subduction (Cluzel & Meffre, 2002). The present location of the exhumed blueschist-facies rocks roughly marks the boundary between Central and Teremba terranes; therefore, the exhumation of the previously subducted material is likely to have taken place within the fore-arc region of the Mesozoic volcanic-arc that generated the volcaniclastic turbidite. Late Jurassic metamorphic ages (ca. 150 Ma, whole-rock K-Ar) of the metabasalt (Blake et al., 1977) are incorrect and probably influenced by excess argon, because detrital zircon ages from metagreywackes at the same locality set a maximum Early Cretaceous (< 130 Ma) depositional age for the original sediments (Adams et al., 2009).

To summarise, the New Caledonia sector of the Gondwanaland margin, which was previously considered as a continental active margin during the pre-Late Cretaceous period (Paris, 1981), can be now interpreted as a forearc sedimentary basin based upon trapped "oceanic" crust (Aitchison et al., 1995; Meffre et al., 1996), fed by an intraoceanic arc located to the west of present-day New Caledonia and now probably buried under younger sediments in the Lord Howe Rise. The paleogeographic features of the three pre-Late Cretaceous terranes suggest that they were formed during an episode of west-dipping subduction. In addition, the faunal endemism shared with New Zealand and usually referred to as the Maori Bioprovince (Wilckens, 1925; Grant-Mackie et al., 1977), and the provenance of most Triassic- Jurassic sediments inferred from detrital zircon age data, all suggest a discontinuous isolation from the main landmass, possibly related to the opening of a marginal basin (Cluzel & Meffre, 2002; Adams et al., 2009).

The Late Cretaceous in New Caledonia

Overlying the three above-mentioned terranes with angular unconformity (Fig. 2), there is a prominent Late Cretaceous (Coniacian to Campanian) (Paris, 1981), volcanosedimentary unit (classically referred to as Formation à charbon), which is composed of fining upwards marine shallow water sandstone, coal-bearing siltstone, tuffs and volcanic rocks that accumulated in a tidal-zone or a deltaic environment. The Late Cretaceous marine siltstones contain endemic faunas (ammonites and inocerams) (see Paris, 1981) which indicate an isolation from Australia, which is confirmed by the local provenance of detrital zircon populations (see below). The pre-Coniacian unconformity post-dates the final amalgamation of the three aforementioned terranes. Exhumation of high-pressure metamorphic rocks of the Boghen Terrane along the Gondwanaland margin thus occurred between the Barremian (ca. 130 Ma; Adams et al., 2009) and the Coniacian (ca. 89 Ma). Mafic and felsic volcanic rocks and tuffs occur near the base of the Formation à charbon (pre-Campanian), the geochemical features of which contrast with that of pre-Late Cretaceous volcanic rocks (see below). U-Pb dating of zircons extracted from a rhyolite flow near the base of the Fm. à charbon (88.4 Ma; Alexander et al., 2010) confirms the Coniacian age of the Late Cretaceous transgression.

Some evidence for syntectonic sedimentation is recognized within unmetamorphosed Late Cretaceous rocks, especially in the Noumea area, that suggests a link with rifting events. The magmatic activity ceased before the Campanian and is thus restricted to the 89-83.5 Ma interval; and note especially that the magmatic activity ceased when the marginal basins opened (Tasman Sea, New Caledonia and South Loyalty basins). Meanwhile, sedimentation evolved from shallow-water marine sandstone and siltstone toward pelagic siliceous pelite (Maastrichtian), and micrite (Palaeocene to Early or mid-Eocene depending upon diachronous Eocene pre-obduction events). This environmental change may be related to the thermal subsidence that followed the marginal rifting and isolation of New Caledonia from continental sources (Cluzel et al., 1994; Aitchison et al, 1995; Cluzel et al., 2001), in particular from the Lord Howe Rise that was above sea level until the Maastrichtian (McDougall and Van Der Lingen, 1974).

In northern New Caledonia, development of the Eocene high-pressure metamorphic complex (Yokoyama et al, 1986; Ghent et al, 1994; Clarke et al, 1997) as a consequence of the subduction of the northern tip of the Norfolk ridge (Aitchison et al, 1995; Cluzel et al., 1994; Cluzel et al., 2001), and its subsequent Late Eocene exhumation (44-34 Ma) (Spandler et al., 2005; Balwin et al., 2007), have dramatically erased most of the primary features of the terrane protoliths. However, the bulk of the Late Cretaceous to mid-Eocene sequence, although metamorphosed to eclogite or blueschist facies, is similar to the unmetamorphosed sequence of the Noumea area; although Late Cretaceous carbonaceous sediments there are considered more distal (Maurizot et al., 1989). In contrast, a series of monotonous uncorrelated metasediments form the bulk of the Mt Panié massif (Fig. 1),this metasedimentary unit is bordered to the south by elements of the Koh Ophiolite (Meffre, 1995) and has been referred to as "undifferentiated Permian to Cretaceous metasediments" (Carroué, 1971; Paris, 1981). As a whole it is correlated with the Koh-Central Terrane (Maurizot et al, 1989).

Paleo-volcanic evidence for a mid-Cretaceous changing geodynamic setting

A detailed study of paleo-volcanic rocks of New Caledonia is beyond the scope of this paper. However, a comparison of selected geochemical features of Permian to Early Cretaceous with Late Cretaceous volcanic rocks allows a mid-Cretaceous change to be established.

Permian to Early Cretaceous volcanic rocks of the Teremba Terrane display a wide compositional range from basalt to rhyolite (data from Adams et al., 2009), and mainly plot on the total alkali vs. silica diagram (Le Bas et al., 1986; Le Maître, 1989) within the field of alkaline series for mafic rocks and sub-alkaline series for felsic ones (Fig 3) (note that major elements features of volcaniclastic sandstones of Teremba and Koh-Central terranes do not allow reliable plotting on such a diagram). Late Cretaceous volcanic rocks of the Formation à charbon similarly display a wide range of compositions (Table 2, Fig 3). Mafic rocks plot within the domain of alkaline and sub-alkaline series as well; in contrast, most felsic rocks plot in the sub-alkaline series domain. Therefore, on the basis of major elements data only, it is difficult to distinguish pre-Late Cretaceous from Late Cretaceous volcanic rocks. However, although based upon a limited number of samples, incompatible trace element ratios (Hf, Ta, and Th) which are known to reflect source signatures (see Pearce, 1983; Wood, 1980 and references herein) show a significant change, appearing in the mid-Cretaceous. On the Hf/3-Th-Ta triangular plot of Wood (1980) (Fig 4), the Permian to Early Cretaceous volcanic and volcaniclastic rocks mainly plot within the field of (intraoceanic) island-arc lavas. These rocks show a low Ta/(Hf/3.+Th) ratio diagnostic of a metasomatised supra-subduction mantle source and a variable Hf/Th ratio (trend "a") indicating minor or negligible input of continental crust-derived material. Thus, the geochemical features of Permian to Early Cretaceous volcanic rocks suggest that they were formed in a volcanic arc resting upon oceanic or intermediate (thinned) continental crust (Meffre, 1995; Adams et al., 2009).

In contrast, Late Cretaceous volcanic rocks show two contrasting trends. The first is a prominent scatter along the Th-Hf/3 side of the triangular plot (trend "b") and thus show variable contamination by continental crust, with a lesser subduction influence marked by higher Ta contents compared to Permian-Early Cretaceous rocks. Second, approximately half of Late Cretaceous volcanic rocks plot within the field of intraplate continental provinces, with some influence of an "active margin" signature (trend "c"). This association of subduction-related with rift-related volcanic rocks is diagnostic of the period that predates the opening of a marginal basin. It is significant that Early Cretaceous volcanic rocks of the Whitsunday Province (Ewart et al., 1992; Bryan et al., 1997), although 45-10 Ma older, show a similar "melange trend," which indicates that the parent magmas result from mixing of active margin and intraplate sources (Fig. 4) and may be similarly interpreted in terms of crustal thinning (see Australian hinterland section below).

The mid-Cretaceous thus marks the transition from the Permian to Early Cretaceous period of "active margin" activity, to the Late Cretaceous marginal rifting. It is worth noting that "onland" rift-related magmatic activity on present-day New Caledonia ends whilst the Campanian-Paleocene New Caledonia Basin and South Loyalty Basin open to the west and to the east of New Caledonia respectively (Cluzel et al., 2001).

U-Pb dating of detrital zircons

Analytical procedures (see Appendix 1)

Early Cretaceous volcaniclastic sandstones in the Koh-Central Terrane

In the Koh-Central terrane, volcaniclastic sandstones generally display a single zircon population, the average age of which is very close to the fossil age (see Adams et al., 2009). Such features are very similar to those of many volcaniclastic sediments in which the youngest detrital zircons are derived from penecontemporaneous volcanic sources.

Two samples from volcaniclastic turbidite outcrops along the Pouembout River (loc. 2 on Fig. 1), previously mapped as Late Jurassic, surprisingly contain late Early Cretaceous (Albian) and early Early Cretaceous (Hauterivian-Barremian) zircons respectively (Table 1; Fig. 5a & b). The samples were collected from a series of fine grained greywacke beds 1 to 10 cm thick, associated with minor argillite seams and rare debris flow conglomerate; they mainly consist of fine grained volcaniclastic material (andesite, clinopyroxene, feldspar) and display geochemical and isotopic features identical to that of other Mesozoic volcaniclastic sandstones of New Caledonia; therefore they are considered to derive from the same volcanic-arc source (Adams et al., 2009, see also Fig. 4). One sample gave a prominent (80%) population of Albian zircons, (108 ± 5 Ma) and a minor one at ca. 123 ± 4 Ma (Adams et al., 2009). The second sample, which displays similar features, has provided a main population (80%) of Barremian-Hauterivian zircons (131 ± 3 Ma) (this study). This record of Early Cretaceous detrital zircons within the Koh-Central terrane significantly lifts the younger age limit of the volcaniclastic sedimentation (e.g., volcanic arc-related) in New Caledonia, which was previously set upon paleontological grounds at ca. 150 Ma (Paris, 1981).

Mid- and Late Cretaceous metasediments in the metamorphic complex of Northern New Caledonia.

Upstream of Paimboas village, on the southeastern flank of Mt Panié Massif, a series of monotonous blueschist facies rocks crops out along the upper reaches of the Diahot river (location 3 on Fig 1). The protoliths of these metamorphic rocks were volcaniclastic sandstone and argillite intruded by scarce meta-dolerite bodies, an association closely similar to that of the Koh-Central Terrane (Maurizot et al, 1989; Meffre, 1995). Three samples have been processed for zircon dating: i) a tectonic boudin with glaucophane, albite and epidote (PAIMB3) enclosed in strongly sheared blueschist facies schist, ii) a meta-sandstone with the same metamorphic association (PAIMB2); and, iii) a meta-conglomerate with a sandstone matrix (PAIMB 1). The boudin (PAIMB3) contains one single population of upper Early Cretaceous zircons (102 ± 3 Ma, Albian) (Fig 5c). Some of the zircons display a small recrystallisation rim dated at 38 ± 3 Ma (Table 3), clearly related to Eocene high-pressure low-temperature metamorphism. It is suggested by analogy to similar unmetamorphosed volcaniclastic sandstones that the single Albian zircon component represents a magmatic event close to the time of sedimentation. Although no geochemical data is yet available on this rock, and in spite of its relatively high metamorphic grade, it clearly appears on the basis of its petrography and the general features of the enclosing series, that it may be correlative to the Koh-Central Terrane greywacke. The meta-sandstone (PAIMB2) contains a few Late Cretaceous zircons (ca. 86±2 Ma, Coniacian-Santonian) and thus may be correlative to the Formation à charbon; it otherwise displays a wider detrital zircon age range, with groups at 93 ± 2 and 115 ± 2 Ma, and only a few are older (Table 2; Fig 7c). The metaconglomerate (PAIMB 1), together with Cretaceous populations similar to PAIMB 2 (85 ± 6 and 99 ± 3 Ma), also contains a wide range of reworked older zircons that include Jurassic, Triassic, Early and mid-Palaeozoic, and Late Proterozoic zircons (Table 2; Fig 7d), many of which would derive directly from a "continental" source; or alternatively, from older, directly underlying greywackes.

Early Cretaceous rocks in the Boghen terrane.

With only a few exceptions, the metamorphosed and highly sheared Boghen terrane, which mainly consists of sliced and broken volcano-sedimentary rocks: pillow lavas, mafic tuffs, sandstone, and very fine grained meta-sediments, has proven to be almost barren of detrital zircon, in spite of extensive sampling by the authors. However, meta-volcaniclastic sandstone, and carbonaceous meta-sandstone that appear rarely within each of the three massifs that form this terrane, have provided abundant detrital zircon populations, the youngest of which set a maximum age constraint on the enclosing meta-sediment. A detailed description of this terrane is beyond the scope of this paper; however, two representative samples will be described here. A typical broken sandstone formation crops out along the reaches of the Komendu River (location 4 on Fig. 1). It is composed of highly disrupted sandstone beds 5 to

10 cm thick, having angular clasts associated with carbonaceous siltstone or argillite. Compared with previous results that set a maximum age of ca. 180 Ma for similar rocks (Cluzel and Meffre, 2002), much younger zircons were extracted from this sandstone (NCAL 32; Table 2). The youngest components fall within the Early Cretaceous (140 ± 5 Ma, Berriasian-Valanginian) (Fig. 6a). The remainder of the zircon population has features similar to that of most sandstone samples of the Boghen terrane (i.e., a major Late Proterozoic - Early Palaeozoic population and a mid-Palaeozoic gap: Cluzel and Meffre, 2002).

Although meta-volcaniclastic sandstones are not rare in the Boghen terrane, they are generally too fine grained for zircon studies. However, one sample from the Faténaoué River (FTN2a; location 5 on Fig. 1) has provided a well defined population (ca. 30%) of Early Cretaceous zircons (133 ± 5 Ma, Hauterivian; Table 1) the youngest grain being 130 ± 3 Ma old (Table 2), together with older Palaeozoic and Early Mesozoic ones (Fig. 6b). From the occurrence of ca. 130 Ma detrital zircon, one can infer a slightly younger depositional age. Therefore, the maximum age of the Boghen terrane is now shown to be 50 Ma younger than previous estimates. It is significant that most greenschists/meta-volcaniclastic sandstones in this terrane display geochemical features of primitive arc-related (island-arc tholeiite) volcaniclastic rocks (Cluzel & Meffre, unpub. data). Therefore, although more investigation is needed, the Fatenaoué metavolcaniclastic sandstone probably signals a prominent phase of previously unrecognized Early Cretaceous volcanic-arc activity.

"Formation à charbon"

An extensive study of detrital zircon provenances in this unit is still in progress; however, preliminary results from sandstone samples from the unconformable cover of the Koh-Central terrane near Koh (loc. 1 on Fig. 1) (Aitchison et al, 1998; Adams et al., 2009) reveal the importance of mid-Cretaceous zircon populations in the "Formation à charbon" and the rarity of older "continent-derived" zircons. Two more samples, a coarse sandstone of the Noumea region (Boulari) (location 7 on Fig. 1) and a slightly metamorphosed sandstone from near the Paimboas village in the north of the island (location 6 on Fig 1) similarly display a very limited range of reworked zircons that include a dominant Aptian-Albian population (Table 2; Fig. 7a & b).

Together with rare Coniacian-Santonian zircons which are probably related to volcanic activity contemporaneous with sedimentation, the occurrence of prominent Early Cretaceous (130-95 Ma) and minor Early to Middle Jurassic (220-160 Ma) zircon components, and also the rarity of older zircons, all suggest that the bulk of the Late Cretaceous sandstone is derived from locally eroded pre-Late Cretaceous terranes. For example, Early Cretaceous zircons may come from "Pouembout-type" volcaniclastic turbidite (greywackes, see above), or from the younger zircon components in the Boghen terrane rocks (Faténaoué-type see above). A close local origin is consistent with the mineral petrography of the studied sample, which contains dominantly angular elements clearly derived from directly underlying rocks; e.g.: clinopyroxene and chert from Koh ophiolite and its abyssal sedimentary cover; schist, epidote and serpentinite clasts from the Boghen terrane; feldspar and some lithic clasts from Mesozoic volcaniclastic sandstone, etc. A local origin for Late Cretaceous sandstone is consistent with the endemic character of the ammonite faunas contained in the sediment.

These results further indicate that Early Cretaceous volcaniclastic sandstones and/or volcanic rocks were much more widespread than previously estimated, and much had probably been eroded before Coniacian time. A minor component of 200-160 Ma Jurassic zircons is also present (Table 1, Fig. 7) and is most likely also derived from eroded volcaniclastic sandstones. It is worth noting that apparent absence of Late Jurassic-Early Cretaceous zircons in these samples was also mentioned by Aitchison et al. (1998).

Implications for New Caledonia's geological evolution

From the evidence presented above, some new constraints on the geological evolution of New Caledonia may be set out as follows:

i) the subduction-related "active margin" evolution of New Caledonia probably paused during the 150-130 Ma interval.

ii) Subduction-related magmatism continued then until mid-Cretaceous time (ca. 95 Ma), instead of 150 Ma as previously postulated.

iii) Most Early Cretaceous volcaniclastic turbidites have been extensively eroded before the Coniacian.

iv) A prominent change occurred in the mid-Cretaceous and an intraplate-type volcanic activity appeared during the Coniacian-Santonian; however, the coexistence of volcanic-arc magmas indicates the persistence of a metasomatised supra-subduction mantle below New Caledonia at that time.

v) Late Cretaceous shallow water sediments (Formation à charbon) mainly contain mid-Cretaceous zircons, probably reworked from the directly underlying "basement" rocks.

Correlative terranes and events in New Zealand

The Mesozoic terranes that form the basement of New Zealand are comprised of mostly Palaeozoic and Mesozoic sedimentary rocks, active-margin volcanic arcs, and associated plutonic complexes. These are divided into an early Paleozoic foreland (Western Province), related to the Lachlan Fold Belt of southeast Australia, and a accretionnary belt (Eastern Province), a collage of several late Paleozoic-Mesozoic tectonostratigraphic terranes. The boundary between the two provinces is occupied by a Median Batholith of Permian-Jurassic plutonic rocks (Bishop et al. 1985, Mortimer et al. 1999). The Median Batholith may be correlated with the more extensive igneous complexes of similar age within the New England Fold Belt of northeast Australia.

The Eastern Province terranes of New Zealand comprise an eastern group (Torlesse, Waipapa and Caples) of Permian to Cretaceous, volcaniclastic sandstone-dominated turbiditic sequences probably deposited on an "oceanic" crust and thereafter accumulated in an accretionary prism environment (Bishop et al. 1985). The remaining, western, terrane group (Dun Mountain-Maitai, Murihiku, Brook Street), has mainly Permian, Triassic and Jurassic, redeposited volcaniclastic sandstone-dominated successions (Ballance & Campbell 1993, Landis et al. 1999), but in a more shallow-water setting (with limestones and also calc-turbidites). The Dun Mountain-Maitai terrane includes the major Dun Mountain Ophiolite Belt (Coombs et al. 1976), an "oceanic" suture (Davey, 2005) which extends through the South and North Islands of New Zealand, and probably farther north (Williams et al, 2006). It represents the remnants of a Permian-Mesozoic back-arc basin (Kimbrough et al., 1992; Sivell & McCulloch, 2000). The relative positions of the terranes, with respect to the Gondwana continental margin, suggest that several, especially the Torlesse composite terrane, have been tectonically displaced from elsewhere along the Gondwana margin. Original sedimentary depocentres have been suggested in the New Zealand-West Antarctic region (Cawood et al. 1999, Wandres et al., 2004a, b); however, Devonian zircons from the extensive Lachlan Fold Belt are very minor in the detrital zircon age patterns, and a required (and extensive) Late Permian-Early Triassic zircon source in this region is absent. Alternatively, original depocentres in the northeastern Australian sector have been suggested (Ireland, 1992; Pickard et al, 2000; Adams et al., 2007), and source rocks in this region would show superior matches of Eastern Province detrital zircon age patterns. In contrast, the Murihiku Terrane which has endemic faunas that closely resemble those of the Teremba Terrane of New Caledonia (Campbell et al. 1985) may be more closely related to New Caledonian terranes.

In the South Island of New Zealand, intrusion of the younger arc-related granitoids and adakitic plutons of the Median Batholith indicates the subduction of relatively young and hot "oceanic" lithosphere (Mortimer et al, 1999). Preserved in the Dun Mountain ophiolitic melange, there are undated, but obviously younger, boninite-like felsic dykes (Sivell & McCulloch, 2000) that crosscut mafic and ultramafic rocks as well, and which may represent fore-arc magmas formed during the early stages of basin closure. Therefore, to account the diverse provenances of deep fan volcaniclastic sandstone series, occurrence of arc-type volcanics and also the biogeographic features (i.e. occurrence of endemic faunas), a geodynamic model including the opening and the subsequent oblique subduction of a marginal basin has been proposed, in which the Dun Mountain ophiolitic melange represents the suture of a Permian-Early Mesozoic obliquely-closed marginal basin, and New Caledonian terranes represent the discontinuously active fore-arc (Adams et al., 2009).

From biostratigraphic evidence in Torlesse Terrane rocks, the active margin activity in New Zealand ceased by mid-Cretaceous time (Motuan stage of New Zealand). It is thought that subduction ceased at ca.100 Ma, when the Hikurangi Plateau, part of the dismembered gigantic Hikurangi-Manihiki-Otong Java plateau (Taylor, 2006), blocked plate margin subduction at the latitude of New Zealand (Davy et al., 2008). However, magmatic activity until 95 Ma is recorded by the occurrence of younger detrital zircons in some volcaniclastic sandstone terranes (Cawood et al, 1999); for example, in the Torlesse Supergroup (Omaio Gr.), the youngest zircons are Albian. The collision timing of Hikurangi Plateau at ca. 100 Ma was predicated on the timing of cessation of spreading along the Osbourn Trough; thus, it is mainly based upon the correlation of magnetic anomalies and a reassessment of the latter might change the interpretation, but not the fact that subduction probably ceased at that time.

Rift-related ductile shear zones that affect Early Cretaceous granitic plutons started at ca. 115 Ma and evolved into half-grabens as late as ca. 88 Ma (K-Ar) (Tulloch & Kimbrough, 1989). In the Otago Schist, age constraints suggest that extensional shear zone formation took place between 135 and 105 Ma, and probably overlaps the timing of marginal rifting (Deckert et al., 2002). It is possible, although not strictly proven, that syn-convergence extensional exhumation at the accretionnary wedge overlapped ductile extension related to the marginal break-off. After the deposition of syntectonic coarse clastic sediments in narrow submarine half-grabens (Tulloch & Kimbrough, 1989), there was a recurrence of Late Coniacian-Santonian (ca. 87-84 Ma) disconformable deposition of coarse, fining-upwards, sandstone (Laird & Bradshaw, 2004). These are overlain by Maastrichtian-Palaeocene hemipelagic deposits, that reflect the initial thermal uplift and later cooling and subsidence of the passive margin, coincident with seafloor spreading in the Tasman Sea. This sequence of events is very similar to that of New Caledonia. The Cretaceous extension-related magmatic activity started in New Zealand and on the Lord How Rise as early as 112 Ma, and continued until 82 Ma (Tulloch et al., 2009). It shows mixed features with typical volcanic-arc signatures and intra-plate signatures as well (Nicholson et al., 2008; Tulloch et al., 2009) (Fig 4).

Australian hinterland in the Early Cretaceous

Marginal rifting in eastern Australia probably started in the Early Cretaceous. The Whitsunday Volcanic Province and Eromanga Basin system, in north-eastern Australia, and the Otway-Gippsland basin system along the south-eastern margin of Australia, are regarded as evidence for this event. The Whitsunday Volcanic Province is part of a high-K calc-alkaline pyroclastic volcanic belt that extends along the central and southern Queensland coast. Unless there is a drastic age revision using new dating methods, these rocks show a rather broad range of ages (K-Ar and fission tracks) from 132 to 95 Ma, but climaxing at 120-105 Ma (Ewart et al, 1992; Bryan et al., 1997). Meanwhile, sedimentary basins in eastern Queensland were receiving large volumes of volcanogenic sediment. The Aptian–Albian Otway and Gippsland Basins mainly consist of intrabasinal volcanic rocks that are probably related to the rifting episode, and younger extrabasinal volcanogenic sediment supplied from the east; i.e. probably from the southern extension of the Whitsunday volcanic belt (Bryan et al., 1997). There is no general agreement on the geodynamic setting of these basins and the associated magmatic activity. Some authors propose a rifting environment which is apparently supported by geochemical evidence from volcanic rocks (Bryan et al., 1997; Bryan, 2005; Fig. 4) whilst others, mainly on the basis of subsidence modelling, favour a back-arc model in which Early Cretaceous basins opened under continuous subduction influence (Waschbusch et al., 2009; Korsch and Totterdell, 2009).

A Whitsunday provenance for some of the Early Cretaceous zircons found in Late Cretaceous sandstone of New Caledonia merits consideration; however, the overall features of the "Formation à charbon", and the lack of older zircon populations of doubtless Australian provenance (e.g. the New England Orogen which is closely associated to Whitsundays volcanic rocks, see Fig 8) do not support such a remote origin.

Discussion: Cretaceous evolution of the Southwest Pacific margin in New Caledonia.

The analysis of detrital zircon age populations of Late Mesozoic volcaniclastic sandstone, metavolcaniclastic sandstone, and "normal" sandstone of New Caledonia suggest almost continuous magmatic activity characterised by subduction-related features in the 250-150 Ma and 130-95 Ma intervals, ca. 10 million years before the opening of Tasman Sea, New Caledonia and South Loyalty basins (Fig. 9). In contrast, extension-related magmatic activity is restricted to a short period ca 89-83 Ma. Deciding whether magmatic activity in eastern Australia or in New Zealand is directly related to an asthenospheric uplift or a back-arc process, is beyond the scope of this paper. However, it is important that the younger subduction-related volcaniclastic sandstones of the Central terrane (ca. 95 Ma) are more or less synchronous with the end of rifting events in eastern Australia, while rift-related magmatic activity in New Caledonia is much younger (< 89 Ma) than in Australia or in New Zealand. Such diachronism is consistent with the general style of Gondwana breakup during which there was progradation southward or south south-eastward in Mesozoic time. Lateral flow of the asthenosphere associated with a ‘scissors-like’ break-off of the continental lithosphere may have had effects similar to an asthenospheric uplift, i.e. generation of silicic large igneous provinces (Bryan et al., 1997; Bryan, 2005), that differ mainly from classical LIP by the involvement of lower continental crust melts.

Biostratigraphic data, and detrital zircon populations of volcaniclastic sandstones also, record a Late Jurassic-Early Cretaceous time-gap, which may be related to a period of relative quiescence of the subduction zone in the vicinity of New Caledonia. In the Boghen terrane, occurrence of Early Cretaceous zircons associated with older "Australian" zircon components records the continuing magmatic activity located farther west. We suggest that the apparent slow-down of paleo-Pacific subduction in the late Jurassic-early Cretaceous was due to the subduction and exhumation of the Boghen accretionary complex (Fig 10a). Thereafter, subduction recommenced for a while and generated the magmatic products that now appear as reworked components in Albian volcaniclastic sandstones and in Late Cretaceous sandstones. Farther south, cessation of subduction at New Zealand latitudes at ca.100 Ma, by the Hikurangi Plateau (Davy et al, 2008) was responsible to the progressive extinction of magmatic activity, because subduction-related volcaniclastic sediments accumulated until ca. 95 Ma in New Zealand and New Caledonia as well.

The intraplate magmatic activity that is closely associated with the opening of continental rifts started as soon as 130-125 Ma in eastern Australia, at ca 100 Ma in New Zealand, and much later, at ca. 89-85 Ma in New Caledonia (Fig 9). Finally, fast eastward roll-back of the Pacific slab, mega-boudinage of the continental lithosphere, and marginal basin formation during the Campanian to Palaeocene, ca. 83-55 Ma, sliced the former margin into elongate submarine rises separated by "oceanic" basins (Fig 10c). Meanwhile, this process also opened such basins (e.g., the South Loyalty Basin; Cluzel et al., 2001) to the east of the Norfolk Ridge, the easternmost known fragment of Gondwanaland. The opening of a marginal basin to the east of the Norfolk ridge would infer the existence of a Late Cretaceous-Paleocene volcanic arc to the east, beyond the South Loyalty Basin. However, there is no trace of such an arc, unless it is buried below younger volcanic-arc rocks in the Fiji-Tonga region; and at this time, no evidence for such older basement rocks has been found. It is possible however that the prominent eastward flow of the upper asthenosphere associated with the initial stretching of the Australian margin and further, the opening of "oceanic" basins of ca. 2,000 km cumulated width, prevented the usual corner flow (uplift of metasomatised asthenosphere controlled by downward drag of the mantle wedge) process to operate, and a volcanic-arc may have never existed at the eastern end of the system. No geochemical data on the rocks that form the oceanic floor of the Tasman Sea and New Caledonia basins is available; however, the "enriched" character of the MORB which compose the obducted part of the South Loyalty Basin (the Poya Terrane of New Caledonia) (Cluzel et al., 2001), suggest the involvement of an asthenospheric mantle component consistent with such an origin.

References

Adams, C. J., Campbell, H. J., & Griffin, W. R. 2007. Provenance Comparisons of Permian to Jurassic Tectonostratigraphic Terranes in New Zealand: Perspectives from detrital zircon age patterns. Geological Magazine 144, 701-729.

Adams C.J., Cluzel D., and Griffin W.L. 2009. Detrital zircon ages and provenance of sedimentary rocks in basement Mesozoic terranes of New Caledonia. Australian Journal of Earth Sciences 56, 1023-1047. DOI: 10.1080/08120090903246162

Aitchison, J. C., Clarke, G. L., Cluzel, D., & Meffre, S. 1995. Eocene arc-continent collision in New Caledonia and implications for regional southwest Pacific tectonic evolution, Geology, 23, 161-164

Aitchison, J. C., Ireland, T. R., Cluzel, D., Davis, A. M., & Meffre, S. 1998. Regional implications of U/Pb SHRIMP age constraints on the tectonic evolution of New Caledonia. Tectonophysics 299, 333-343.

Alexander, A. M., Nicholson, K. N., Simonetti, A., Maurizot, P., and Cluzel, D., 2010. Dating rhyolite flows in the Noumea basin, New Caledonia, in order to constrain the breakup of the eastern Gondwana margin. GSA 44th Annual Meeting, Branson, Mi, USA. April 11-13, 2010. Abstracts with Programs V. 42, n° 2.

Andersen, T. 2002. Correction of common Pb in U-Pb analyses that do not report 204Pb. Chemical Geology (Isotope Geoscience) 192, 59-79.

Andersen, T. 2005. Detrital zircons as tracers of sedimentary provenance: limiting conditions from statistics and numerical simulation. Chemical Geology (Isotope Geoscience) 216, 249-270.

Ballance, P. F., & Campbell, J. D. 1993. The Murihiku arc-related basin of New Zealand (Triassic-Jurassic). *In*: Ballance, P.F. (ed). South Pacific Sedimentary Basins, Sedimentary Basins of the World. Elsevier, Amsterdam, 21-33.

Betts P.G., Giles D., Lister G. S., and Frick L. R. 2002. Evolution of the Australian lithosphere. Australian Journal of Earth Sciences 49, 661-695.

Bishop, D. G., Bradshaw, J. D. & Landis, C. A. 1985. Provisional terrane map of South Island, New Zealand. *In*: Howell, D.G. (ed) Tectonostratigraphic Terranes, Circum - Pacific Council for Energy and Mineral Resources Earth Science Series No. 1, Houston, Texas, 515-521.

Black, L.P. et al., 2003. TEMORA 1: a new zircon standard for Phanerozoic U–Pb geochronology. Chemical Geology, 200: 155– 170.

Blake, M. C., Brothers, R. N., & Lanphere, M. A. 1977. Radiometric ages of blueschists in New Caledonia. In International Symposium on Geodynamics in the South West Pacific. Nouméa, 1976, Edition Technip, Paris, 276-282.

Bradshaw J.D. 1989. Cretaceous geotectonic patterns in the New Zealand region. Tectonics 8, 803-820.

Bryan S.E., A.E. Constantine, C.J. Stephens, A. Ewart, R.W. Schon, J. Parianos, 1997. Early Cretaceous volcano-sedimentary successions along the eastern Australian continental margin: Implications for the break-up of eastern Gondwana. Earth and Planetary Science Letters 153, 85-102

Bryan S.E. 2005. The Early Cretaceous Whitsunday Silicic Large Igneous Province of Eastern Australia. Large Igneous Province Commission, I.A.V.C.E.I., LIP of the month August 2005. http://www.largeingneousprovinces.org

Campbell, H. J. 1984. Petrography and metamorphism of the Téremba Group (Permian to Lower Triassic) and the Baie de St. Vincent Group (Upper Triassic to Lower Jurassic), New Caledonia. Journal of the Royal Society of New Zealand, 14, 335-348.

Campbell, H. J., Grant-Mackie, J. A., & Paris, J. P. 1985. Geology of the Moindou-Téremba area, New Caledonia; stratigraphy and structure of the Teremba Group (Permian-Lower Triassic) and Baie de St. Vincent Group (Upper Triassic-Lower Triassic). Géologie de la France, BRGM, 1, 19-36.

Carroué, J. P., 1971, Carte géologique de la Nouvelle-Calédonie à l'échelle du 1/50 000, feuille Pouébo, B.R.G.M., Paris, map sheet and explanatory notes, 12 p.

Cawood, P.A., Nemchin, A.A., Leverenz, A., Saeed, A., Ballance, P.F. 1999. U/Pb dating of detrital zircons : Implications for the provenance record of Gondwana margin terranes. Geological Society of America Bulletion 111, 1107-1119.

Clarke G., J.C. Aitchison and D. Cluzel, 1997. Eclogites and blueschists of the Pam Peninsula, NE New Caledonia : a reappraisal. Jour Metam. Petrology 38, 7, 843-876.

Cluzel, D., Aitchison, J., Clarke, G., Meffre, S., and Picard, C., 1994. Point de vue sur l'évolution tectonique et géodynamique de la Nouvelle-Calédonie, C. R. Acad Sci. Paris. 319, 683-688.

Cluzel D., Aitchison J.C., Black P.M. and Picard C., 1999. Origin and fate of Southwest Pacific marginal basins; an appraisal from New Caledonia. Geological Society of America, ‘99 Penrose Conference, Arthur’s Pass, South Island, New Zealand, Abstr. vol., 21-22.

Cluzel, D., Aitchison, J.C., and Picard, C., 2001. Tectonic accretion and underplating of mafic terranes in the Late Eocene intraoceanic fore-arc of New Caledonia (Southwest Pacific): geodynamic implications. Tectonophysics 340, 23-59.

Cluzel, D., and Meffre, S. 2002. L'unité de la Boghen (Nouvelle-Calédonie, Pacifique sud-ouest): un complexe d'accrétion jurassique. Données radiochronologiques préliminaires U-Pb sur les zircons détritiques. C.R. Géosciences 334, 867-874.

Cluzel D., Black P., Picard C., and Nicholson K., 2010. Geochemistry and tectonic setting of Matakaoa Volcanics (East Coast Allochthon, New Zealand); supra-subduction zone affinity, regional correlations and origin. Tectonics. (in press) doi:10.1029/2009TC002454.

Cluzel D., and Meffre S. (in prep.). Sedimentary and magmatic heritage of the Boghen Metamorphic Terrane (New Caledonia, Southwest Pacific), a Late Mesozoic accretionnary complex at the Southeast-Gondwana active margin.

Coombs, D.S., Landis, C.A., Norris, R.J., Sinton, J.M., Borns, D.J., and Craw, D. 1976. The Dun Mountain Ophiolite, New Zealand, its tectonic setting, constitution, and origin, with special reference to the southern portion. American Journal of Science 276, 561-603.

Cooper R.A. (compiler) 2004. New Zealand Geological Timescale 2004/2 wallchart. Institute of Geological & Nuclear Sciences information series 64.

Courtillot V., Jaupart C., Manighetti I., Tapponnier P. and Besse J. 1999. On causal links between flood basalts and continental breakup. Earth and Planetary Science Letters166, 3-4, 177-195; doi:10.1016/S0012-821X(98)00282-9

Davey F.J. 2005. A Mesozoic crustal suture on the Gondwana margin in the New Zealand region. Tectonics 24, TC4006, doi:10.1029/2004TC001719.

Davy B., Hoernle K. and Werner R. 2008. Hikurangi Plateau: Crustal structure, rifted formation, and Gondwana subduction history. Geochemistry, Geophysics, Geosystems 9, Q07004, doi:10.1029/2007GC001855

Deckert H., Ring U. and Mortimer N. 2002. Tectonic significance of Cretaceous bivergent extensional shear zones in the Torlesse accretionnary wedge, central Otago Schist, New Zealand. New Zealand Journal of Geology and Geophysics 45, 537-547.

Eagles G. and König M., 2008. A model of plate kinematics in Gondwana breakup Geophys. J. Int., 173, 703-717; doi: 10.1111/j.1365-246X.2008.03753.x

Ewart A., Schon R.W., Chappell B.W., 1992. The Cretaceous volcanic–plutonic province of the central Queensland Australia.coast - a rift related "calc-alkaline" province, Trans. R. Soc. Edinburgh Earth Sci. 83, 327-345.

Gaina C., Roest W.R., Müller R. D., and Symonds P., 1998a. The Opening of the Tasman Sea: A Gravity Anomaly Animation. Earth Interactions, 2, 1-23.

Gaina, C., Müller R.D., Royer J.-Y., Stock J., Hardebeck J., and Symonds P., 1998b. The tectonic evolution of the Tasman Sea: A tectonic puzzle with thirteen pieces. J. Geophys. Res., 103, (B6), 12,413–12,433.

Ghent, E.D., Roddick, J.C., Black, P.M., 1994. $^{40}Ar/^{39}Ar$ dating of white micas from the epidote to omphacite zone, northern New Caledonia : tectonic implications, Can Jour. Earth Sci, 31, 995-1001

Grant-Mackie, J. A., Paris, J. P. Freneix, S., & Campbell, J. D. 1977. Advances in correlation of Mesozoic sequences in New Zealand and New Caledonia. 25th International Geological Congress, Abstracts, 1, 268-269.

Guérangé, B., Lozes J., Autran A., 1977. Le métamorphisme mésozoïque dans la chaîne centrale de Nouvelle-Calédonie et ses implications géodynamiques dans l'évolution de l'orogenèse Rangitata au Crétacé. Bull. BRGM Fr., sect. IV, 1, 53-68.

Ireland, T.R. 1992. Crustal evolution of New Zealand: Evidence from age distributions of zircons in Western Province paragneisses and Torlesse greywacke. Geochimica et Cosmochimica Acta 56, 911-920.

Irvine, T. N. & Baragar, W. R. A. 1971. A guide to the chemical classification of the common volcanic rocks. Canadian Journal of Earth Sciences 8, 523–548.

Jackson, S. E., Pearson, N. J., Griffin, W. L., & Belousava, E. A. 2004. The application of laser abalation-inductively coupled plasma-mass spectrometry (LA-ICP-MS) to in situ U-Pb zircon geochronology. Chemical Geology, 211, 47-69.

Kimbrough, D. L., Mattinson, J. M., Coombs, D. S., Landis, C. A., & Johnston, M. R. 1992. Uranium-lead ages from the Dun Mountain Ophiolite Belt and Brook Street Terranes, South Island, New Zealand. Geological Society of America Bulletin 104, 429-443.

Korsch R. J., and Totterdell J. M. 2009. Evolution of the Bowen, Gunnedah and Surat Basins, eastern Australia. Australian Journal of Earth Sciences 56, 3, 271-272.

Laird M.G., and J.D. Bradshaw 2004. The Break-up of a Long-term Relationship: the Cretaceous Separation of New Zealand from Gondwana. Gondwana Research 7, 273-286

Landis, C.A., Campbell, H.J., Aslund, T., Cawood, P.A., Pillai, D.D.L., Raine, J.I., Willsman, A. 1999. Permian-Jurassic strata at Productus Creek, Southland, New Zealand: implications for terrane dynamics at the eastern Gondwanaland margin. New Zealand Journal of Geology and Geophysics 42, 255-278.

Lister G.S., and Etheridge M.A., 1989. Detachment models for uplift and volcanism in the Eastern Highlands, and their implication to the origin of passive margin mountains, in: R.W. Johnson, S.R. Taylor Eds., Intraplate Volcanism in Eastern Australia and New Zealand, Cambridge University Press, 297–313.

Luyendyk B. P. 1995. Hypothesis for Cretaceous rifting of east Gondwana caused by subducted slab capture. Geology 23; 4; 373–376.

Le Bas, M. J., Le Maitre, R. W., Streckeisen, A. & Zanettin, B. 1986. A chemical classification of volcanic rocks based on the total alkali–silica diagram. Journal of Petrology 27, 745–750.

Le Maitre, R. W. 1989. A Classification of Igneous Rocks and Glossary of Terms. Recommendations of the IUGS Commission on the Systematics of Igneous Rocks. Oxford: Blackwell.

Lister G.S., and Etheridge M.A., 1989. Intraplate volcanism in eastern Australia and New Zealand, 1989 - Cambridge University Press

Luyendik B.P. 1995. Hypothesis for Cretaceous rifting of east Gondwana caused by subducted slab capture. Geology 23, 373-376.

MacKinnon, T. C. 1983. Origin of the Torlesse Terrane and coeval rocks, South Island, New Zealand. Geological Society of America Bulletin 94, 967-985.

Marzoli A., Renne P.R., Piccirillo E.M., Ernesto M., Bellieni G., De Min A. 1999. Extensive 200-Million-Year-Old Continental Flood Basalts of the Central Atlantic Magmatic Province *Science* 284, 5414, 616 – 618; DOI: 10.1126/science.284.5414.616

Maurizot , P., Eberlé, J.M., Habault, C., Tessarolo, C., 1989. Carte géol. Territoires d'Outre-Mer, Nouvelle-Calédonie (1/50000), feuille Pam-Ouégoa, 2e édition, B.R.G.M., Notice explicative par Maurizot P., Eberlé J.M., Habault C., et Tessarolo C., 81p.

McDougall, I., & Van Der Lingen, G. J. 1974. Age of the rhyolites of the Lord Howe Rise and the evolution of the Southwest Pacific Ocean. Earth and Planetary Science Letters 21, 117-126.

Meffre, S., 1995. The developpement of arc-related ophiolites and sedimentary sequences in New Caledonia. PhD Thesis, University of Sydney, 236 p.

Meffre, S., Aitchison, J. C., & Crawford, A. J. 1996. Geochemical stratigraphy of boninites and tholeiites from the Permo-Triassic Koh Ophiolite, New Caledonia. Tectonics 15, 67-83.

Mortimer, N., Tulloch, A. J., Gans, P., Calvert & Walker, N. 1999. Geology and thermochronometry of the east edge of the Median Tectonic Batholith (Median Tectonic

Zone): a new perspective on Permian to Cretaceous crustal growth of New Zealand. Island Arc 8, 404-425.

Nicholson, K.N., Black, P.M. and Spörli K.B., 2008. Cretaceous–Oligocene multiphase magmatism on Three Kings Islands, northern New Zealand. New Zealand Journal of Geology & Geophysics 51, 219-229.

Paris, J. P. 1981. La Géologie de la Nouvelle-Calédonie, un essai de synthèse. Mémoire BRGM, 133, 250pp.

Pearce J. A. 1983. Role of the sub-continental lithosphere in magma genesis at active continental margins. In: Hawkesworth C. J. & Norry M. J. eds. Continental Basalts and Mantle Xenoliths, 230–249. Shiva, Nantwich.

Picard, M., 1995. Ouverture de bassins marginaux dans le Pacifique sud-ouest. Paléo-volcanisme et reconstitution des environnements géodynamiques. Extrémité nord de la Nouvelle-Calédonie, Mém. DEA Géodynamique des marges océaniques et des chaînes de montagnes, Univ. Grenoble, 124 p.

Pickard, A. L., Adams, C. J., & Barley, M. E. 2000. Australian provenances for Upper Permian to Cretaceous rocks forming accretionary complexes on the New Zealand sector of the Gondwanaland margin. Australian Journal of Earth Sciences 47, 987-100.

Schellart W.P., Lister G.S., Toy V.G. 2006. A Late Cretaceous and Cenozoic reconstruction of the Southwest Pacific region: Tectonics controlled by subduction and slab rollback processes. Earth Science Reviews 76, 191-233.

Sivell, W.J. & Mc Culloch, M.T. 2000. Reassessment of the origin of the Dun Mountain Ophiolite, New Zealand: Nd-isotopic and geochemical evolution of magma suites. New Zealand Journal of Geology & Geophysics 43, 133-146.

Spörli, K.B.1978. Mesozoic tectonics, North Island, New Zealand. Geological Society of America Bulletin 89, 415-425.

Storey B.C., Curtis M.L., Ferris J.K., Hunter M.A. and Livermore R.A., 1999. Reconstruction and break-out model for the Falkland Islands within Gondwana. Journal of African Earth Sciences. Vol. 29, No. 1, PP. 153-l 63.

Taylor B. 2006. The single largest oceanic plateau: Ontong Java–Manihiki–Hikurangi. Earth and Planet. Sci. Let. 241, 372-380.

Tulloch, A. J., and D. L. Kimbrough (1989), the Paparoa metamorphic core complex, New Zealand: Cretaceous extension associated with fragmentation of the Pacific margin of Gondwana, Tectonics, 8(6), 1217–1234.

Tulloch A. J., Ramezani J., Mortimer N., Mortensen J., van den Bogaard P., & Maas R. 2009. Cretaceous felsic volcanism in New Zealand and Lord Howe Rise (Zealandia) as a precursor to final Gondwana break-up. Geological Society, London, Special Publications; 321, 89-118; DOI: 10.1144/SP321.5

Turnbull, I.M. 1979. Stratigraphy and sedimentology of the Caples terrane of the Thomson Mountains, northern Southland, New Zealand. New Zealand Journal of Geology and Geophysics 22, 555-574.

Veevers J.J., Powell C.M.A., and Roots S.R. 1991. Review of seafloor spreading around Australia. I. Synthesis of the patterns of spreading. Australian Journal of Earth Sciences 38, 4, 373-389.

Veevers J. J. 2000. Change of tectono-stratigraphic regime in the Australian plate during the 99 Ma (mid-Cretaceous) and 43 Ma (mid-Eocene) swerves of the Pacific. Geology; 28; 47-50; DOI: 10.1130/0091-7613(2000)28<47:COTRIT>2.0.CO;2

Veevers J.J. 2004. Gondwanaland from 650–500 Ma assembly through 320 Ma merger in Pangea to 185–100 Ma breakup: supercontinental tectonics via stratigraphy and radiometric dating. Earth Science Reviews 68, 1-132.

Walker, J.D., and Geissman, J.W. (compilers) 2009. Geologic Time Scale: Geological Society of America, doi: 10.1130/2009.CTS004R2C.

Wandres, A. M., Bradshaw, J. D., Weaver, S., Maas, R., Ireland, T. R, & Eby, N. 2004a. Provenance analysis using conglomerate clast lithologies: a case study from the Pahau terrane of New Zealand. Sedimentology 167, 57-89.

Wandres, A. M., Bradshaw, J. D., Weaver, S., Maas, R., Ireland, T. R., and Eby, N. 2004b. Provenance of the sedimentary Rakaia sub-terrane, South Island, New Zealand: the use of igneous clast compositions to define the source. Sedimentology 168, 193-226.

Waschbusch P., Korsch R. J., Beaumont C. 2009. Geodynamic modelling of aspects of the Bowen, Gunnedah, Surat and Eromanga Basins from the perspective of convergent margin processes. Australian Journal of Earth Sciences, 56, 3, 309-334.

Wilckens O. 1925. Stratigraphie und bau von Neu Caledonien. Geol. Rdsch. Leipzig, 16, 2, 128-142.

Williams, H.A., Cassidy, J., Corrine, A., Locke, K. & Spörli, B. 2006. Delineation of a large ultramafic massif embedded within a major SW Pacific suture using gravity methods. Tectonophysics 424, 119-133.

Wilson M. 1997. Thermal evolution of the Central Atlantic passive margins: continental break-up above a Mesozoic super-plume. Journal of the Geological Society, 154; 3; 491-495; DOI: 10.1144/gsjgs.154.3.0491

Wood, D. A.1980. The application of a Th-Hf-Ta diagram to problems of tectonomagmatic classification and to establishing the nature of crustal contamination of basaltic lavas of the British Tertiary Volcanic Province. Earth and Planetary Science Letters, 50, 1, 11-30. DOI: 10.1016/0012-821X(80)90116-8

Yokoyama K., R.N. Brothers, and P.M. Black, 1986. Regional eclogite facies in the high-pressure metamorphic belt of New Caledonia. In: B.W. Evans and E.H. Brown, Editors, Blueschists and Eclogites, Mem. Geol. Soc. Amer., pp. 407–423.

Figure captions

Figure 1: Sketch map of New Caledonia with location of sampling localities referred to in the text (see Table 1). Insert legend: QP: Queensland Plateau; MP: Marion Plateau; DR: Dampier Ridge; LHR: Lord Howe Rise; NR: Norfolk Ridge; LR: Loyalty Ridge; ChP: Challenger Plateau; HP: Hikurangi Plateau; CR: Chatham Rise; CP: Campbell Plateau.

Figure 2: Very simplified cross section of New Caledonia (La Foa – Canala; see location on Fig. 1) to show the three pre-Late Cretaceous terranes tectonically juxtaposed, eroded, and overlain by the unconformable "Formation à charbon".

Figure 3: Total alkali vs. silica diagram after Le Bas et al (1986) and Le Maitre (1989) to compare the major elements compositions of Permian to Early Cretaceous volcanic rocks of the Teremba Terrane (3a) to the Late cretaceous volcanic rocks of the Nouméa region and Diahot Terrane (3b).The dividing line between alkaline and sub-alkaline series is from Irvine & Baragar (1971).

Figure 4: Hf/3 – Th – Ta triangular plot of Wood (1983) to show the distribution of Permian to Early Cretaceous (290-95 Ma) volcanic rocks and greywackes of New Caledonia in the field of oceanic arcs (left triangle; Adams et al., 2009), the star represents the sample NCAL 10 (Albian); and in contrast, the more scattered distribution of Late Cretaceous (ca. 89-83.5 Ma; Table 2) volcanic rocks (right triangle) in the field of active margins and volcanic arcs,

and in the domain of intraplate continental provinces respectively. The fields of Early Cretaceous (130-95 Ma) Whitsunday volcanic rocks (Ewart et al., 1992 in: Bryan et al., 1997), mid-Cretaceous (112-82 Ma) volcanic rocks of New Zealand – Lord Howe Rise (Tulloch et al., 2009), and Late Cretaceous rocks of the Three Kings Islands (New Zealand; Nicholson et al., 2008) are shown for comparison.

Figure 5: Probability density diagram for detrital zircon age populations from Early Cretaceous greywackes of the Koh-Central terrane, 5a & b: Pouembout river, 5c: Paimboas, Diahot river.

Figure 6: Probability density diagram for detrital zircon age populations from the Boghen Terrane. 6a: Faténaoué river, 6b: Komendu river.

Figure 7: Probability density diagram for detrital zircon age populations from the Late Cretaceous "Formation à charbon". 7a: sandstone Ouéhole village (north); 7b: sandstone Boulari (SE of Noumea); 7c: meta-sandstone Paimboas (Diahot); 7d: meta-conglomerate Paimboas (Diahot).

Figure 8. Configuration of the south-eastern Gondwanaland margin before the mid-Cretaceous marginal rifting (modified from Gaina et al. 1998; Sutherland 1999; and Hall 2002) to show the location of major provinces and possible sediment sources cited in the text. The thick black line represents the Dun Mountain Ophiolite Belt. There is still some uncertainty on the precise limits of continental fragments due to the Cretaceous stretching of continental crust.

Figure 9: A summary of event chronology in the Australian margin during the Late Jurassic Oligocene period to show the timing of subduction-related magmatism, eastward migration of rift-related magmatic activity; and in contrast, the simultaneous opening of Tasman, New Caledonia and South Loyalty marginal basins.

Figure 10: Reconstruction of the Jurassic to Palaeocene evolution of the south-eastern Australian margin at the latitude of New Caledonia. 10a: Late Jurassic to Early Cretaceous final closure of the marginal basin(s), accretion of Eastern Province terranes, and formation of the Median Batholith in part (NZ); meanwhile, sediments coming from Australia could not

reach New Caledonia unless following a northern route and were accreted to form the Boghen Terrane; 10b: mid-Cretaceous incipient marginal rifting of south-eastern Gondwanaland and development of the Whitsunday Volcanic Province and possibly of the youngest components of the Medium Batholith; 10c: Latest Cretaceous break-up, large scale boudinage of the lithosphere and "Pacific" slab rollback accompanied by prominent eastward flow of the asthenosphere (10c adapted from Schellart et al., 2006).

Appendix I: TECHNICAL DETAILS

Stages names and limits used in the text refer to the New Zealand timescale of Cooper (2004) and G.S.A. geological timescale of Walker and Geissman (2009).

Zircon dating at GEMOC (CJA)

Samples NCAL 10, NCAL 24, and NCAL 32 were taken at stratigraphic horizons where coarse-medium greywacke and sandstone predominate (Table 1). To minimise sample handling for zircon recovery, a 1 kg sample was collected at the field outcrop as 5 mm-size gravel, removing all weathered rinds, blemishes, inclusions and joint faces. This enabled direct crushing in a tungsten carbide swingmill 2-3 times, for 5-10 seconds, sieving at each stage through a single, 250 micron mesh sieve. The sieved material was washed and decanted several times in water, to remove mud-size fractions, thus retaining a 200-300 g sample in a ~30-250 micron size range, which was then dried. A heavy mineral concentrate was obtained from a 100 g portion in sodium polytungstate liquid, adjusted to a specific gravity of 2.95-2.98, from which about 500 zircon grains were then hand-picked as randomly as possible, i.e. taking all grains within a 1 mm microscope stage field of view. Of these, 50-100 grains were mounted in resin to be polished for LA-ICPMS (laser-ablation inductively-coupled plasma-source mass spectrometry) analysis.

Analytical protocols relating to ablation procedures, mass spectrometric analysis and data treatment are discussed in detail in Jackson et al. (2004). These authors' preferred procedures were followed in this work, using a Merchantek pulsed Nd-YAG laser, frequency-quintupled to operate at 213 nm, and an Agilent 7500S ICPMS instrument.

In all cases, the ablated spot size was in the range 30-50 microns, with the ablation time about 60 seconds, preceded by 60 seconds background measurement, and followed by 60-120 seconds washout. Groups of 10-12 zircon sample grain analyses were preceded and followed by duplicate analyses of firstly, the in-house zircon standard GJ-1, and secondly, by 1-2 analyses each of the international zircon standards, MT-1 and 91500. The GLITTER data interpretation software package (www.els.mq.edu.au/GEMOC/) enabled analysis of U, Pb and Th absolute count rates, and all relevant isotopic ratios, during the run cycle, and the elimination of unstable beam intervals, and rejection of data where zircon core regions were inadvertently encountered.

Using the laser spot size of 30-50 microns enabled age measurements to be made adjacent to crystal margins, rather than cores, and preferably, close to crystal terminations (as defined by two crystal edges). Isotopic data were continually monitored during ablation to check that zircon cores were not being intersected. Efficient use of the instrument time dictated that strongly unimodal patterns were investigated only to analysis totals of N=33-50, bimodal patterns to N=50-70, and strongly polymodal patterns to N=100 (N.B. throughout this work ‘N’ and ‘n’ refer to dataset totals and subgroups respectively). This allowed significant age groups (n) comprising >10% to be revealed by three or more analyses (Andersen, 2005).

Full $^{207}Pb/^{206}Pb$, $^{206}Pb/^{238}U$, $^{207}Pb/^{235}U$, and $^{208}Pb/^{232}Th$ age data (and 1 standard errors) are listed in the Appendix. All ages used here are $^{206}Pb/^{238}U$ zircon ages where <1000 Ma, and $^{207}Pb/^{206}Pb$ ages where >1000 Ma. A small minority of the analyses have common Pb corrections (using protocols of Andersen, 2002). Age groupings, were determined by visual inspection of probability density plots of zircon age sets, using deconvolution (and weighted average) algorithms in the ISOPLOT-Ex (version 3.0) software (kindly provided by K. Ludwig, United States Geological Survey).

Zircon dating at Hobart (SM)

Zircon crystals from samples PAIMB 1, 2, 3, 4, PM 118, NOU 121 and FTN2A were dated using a Hewlett Packard HP4500 ICP-MS fitted with a Merchantek Nd-YAG laser operating at 213 nm at the University of Hobart (Tasmania). The crystals were separated using a gold pan and a magnet, mounted in epoxy blocks, and 30 µm spots on each crystal was sampled by the laser in a He atmosphere. Mass bias, machine drift and fractionation were corrected by analysing the Temora zircon standards of Black et al. (2003). The 1063 Ma 91500 zircons (Wiedenbeck et al., 1995), and an in-house secondary standard widely used by the Australian National University (the 42.2 Ma 98-521 zircons; C. Allen written communication 2004), were analysed in the same analytical runs as the granitoid zircons and gave results within analytical error of the recommended values.

Table 2. U-Pb detrital zircon age data (see Excel file).

Table 1: New Caledonia Cretaceous greywacke and sandstone sample description, location, and data.

Sample	Loc. on Fig 1	location	coordinates	terrane	lithology	assumed age	youngest zircon	youngest zircon population	older populations
NCAL 24	1	near Koh village	21°33'48"S 165°49'57"E	Fm à charbon	sandstone	Coniacian-Santonian	80 ± 2 Ma	85 ± 2 Ma (n=5)	130-95 Ma 200-160 Ma
PAIMB 4	6	south of Paimboas	20°32'47"S 164°30'40"E	Diahot	sandstone	Late Cretaceous	92 ± 5 Ma	103.0 ± 2.2 Ma (n=24)	172 942, 614
NOU 121	7	Boulari	22°13'50"S 166°31'20"E	Fm à charbon	feldspathic sandstone	Coniacian-Santonian	96 ± 2 Ma	101.7 ± 1.5 Ma (n=25)	115, 135 211-189 Ma
NCAL 10	5	Pouembout riv.	21°06'41"S 165°01'25"E	Koh-Central	greywacke	correlated to the Late Jurassic	103 ± 1 Ma	108 ± 2 Ma (n=16)	123 ± 4 Ma (n=5)
PM 118	5	Pouembout riv.	21°06'41"S 165°01'25"E	Koh-Central	greywacke	correlated to the Late Jurassic	121 ± 3 Ma	131 ± 1.2 Ma (n=35)	195-167 Ma 775-500 Ma
PAIMB 3	3	Témélin (Paimboas)	20°30'40"S 164°34'32"E	Koh-Central within Diahot ?	mafic boudin	? (Tertiary HP-LT complex)	94 ± 4 Ma	95.5 ± 1.5 Ma (n=8)	101.6 ± 2 Ma (n=8)
PAIMB 2	3	Témélin (Paimboas)	20°30'40"S 164°34'24"E	Diahot ?	meta-sandstone	? (Tertiary HP-LT complex)	86 ± 2 Ma	93 ± 2 Ma (n=11)	115 ± 4 Ma 180 Ma
PAIMB 1	3	Témélin (Paimboas)	20°32'06"S 164°33'40"E	Diahot ?	meta-conglomerate	? (Tertiary HP-LT complex)	85 ± 6 Ma	99 ± 3 Ma (n=5)	159 ± 5, 250-212, 850-650
FTN2a	2	Faténaoué riv.	20°50'00"S 164°47'41"E	Boghen	metagreywacke	? (Late Mesozoic HP-LT complex)	128 ± 2 Ma	133 ± 5 Ma (n=5)	250-220 Ma 330 Ma 600-450 Ma
NCAL 32	4	Komendu riv.	21°19'26"S 165°24'40"E	Boghen	carbonaceous sandstone	? (Late Mesozoic HP-LT complex)	140 Ma	140 ± 5 Ma (n=3)	170-150 Ma 215 Ma 650-415 Ma

Table 2 Whole-rock major and trace element geochemistry of Late Cretaceous volcanic rocks of the Diahot Terrane and Nouméa area

sample	BP58-1	BP46-3	BP46-2	BP46-1	BP73	BP72	BP71	BP70	BP61	BP60	BP56	BP43	BP24	BP16
lithol	rhyol.	basalt	basalt	basalt	metabas.	metabas.	metabas.	metabas.	metabas.	metabas.	metabas	rhyol.	doler	andesit
Loc.	Fern hill	Bambou	Bambou	Bambou	Ouégoa	Ouégoa	Ouégoa	Ouégoa	Bondé	Bondé		Ougne	Crossroad Mérét	Arama
Terrane	Diahot	Diahot	Diahot	Diahot	Diahot	Diahot	Diahot	Diahot	Diahot	Diahot	Diahot	Diahot	Diahot	Diahot
SiO2	76.22	50.26	46.91	46.63	46.98	50.19	45.00	43.70	45.11	43.09	44.78	74.11	49.15	48.45
A1203	10.78	14.26	14.75	14.74	14.00	13.46	18.22	17.21	17.05	17.20	16.41	11.50	14.50	17.67
Fe203	1.31	10.35	12.21	13.28	11.93	13.63	8.58	9.05	9.83	9.12	9.32	0.75	11.10	9.45
MnO	nd	0.07	0.17	0.17	0.18	0.18	0.11	0.12	0.15	0.14	0.14	nd	0.15	0.13
MgO	0.05	6.59	7.58	5.90	8.47	6.92	5.69	10.77	8.10	9.96	9.83	0.46	7.17	7.23
CaO	nd	8.37	5.11	6.19	10.50	9.86	10.52	9.69	10.27	11.08	10.48	9.21	7.14	4.02
Na20	0.59	3.54	3.81	4.10	2.76	2.96	3.35	2.05	2.46	1.58	2.31	0.09	3.73	4.76
K20	8.74	0.34	1.73	1.48	0.02	0.01	0.02	0.26	0.28	0.26	0.04	0.20	0.51	0.03
TiO2	0.13	1.85	2.49	2.95	1.30	1.77	1.21	0.51	0.94	0.85	0.78	0.14	1.84	1.64
P205	nd	0.26	0.54	0.56	0.17	0.20	0.18	0.09	0.11	0.12	0.12	0.04	0.27	0.34
LOI	0.74	3.93	4.52	3.82	3.51	0.70	6.92	6.40	5.55	6.41	5.76	3.31	4.25	6.18
Total	98.56	99.82	99.82	99.82	99.82	99.87	99.80	99.85	99.85	99.81	99.97	99.81	99.81	99.90
As	12.10	0.48	0.42	0.71	0.23	10.90	2.61	0.13	0.07	0.28	0.38	0.01	1.33	0.02
Ba	729	86.6	224	136	19.70	6.94	24.3	99.4	151	176	13.8	419	98.6	66.4
Be	0.39	1.26	1.46	0.60	nd	nd	nd	nd	0.38	0.23	0.15	nd	nd	0.31
Bi	0.28	nd	nd	0.01	nd	nd	nd	nd	nd	nd	nd	nd	0.02	nd
Cd	0.25	0.22	0.80	0.03	0.12	0.25	1.45	0.13	0.16	0.15	0.14	0.13	0.34	0.11
Ce	21.52	20.91	51.06	61.64	11.62	11.00	13.00	4.27	7.01	5.77	5.96	17.25	26.72	27.04
Co	1.52	33.60	35.10	36.00	46.70	49.80	29.70	41.40	41.10	43.90	42.80	1.96	32.00	28.70
Cr	37.3	371	93.1	16.4	357	138	210	460	304	441	336	51.9	144	204
Cs	0.98	0.68	0.75	0.81	0.20	0.21	0.21	0.46	4.10	5.05	0.33	0.18	0.96	0.21
Cu	23.2	44.9	96.9	50.5	150	79	58.6	53.6	81.4	86.7	70.1	3.80	52.9	60.5
Dy	2.41	4.02	5.21	5.90	3.96	6.64	4.78	2.33	4.03	2.97	3.25	2.11	7.07	7.33
Er	1.66	1.89	2.54	2.78	2.34	4.17	2.89	1.62	2.49	1.89	2.07	1.26	4.26	4.32
Eu	0.42	1.50	2.25	2.52	1.13	1.34	1.21	0.57	0.96	0.86	0.73	0.27	1.62	1.74
Ga	10.4	23.5	24.7	26.4	15.6	19.4	17.2	14.2	16.3	15.1	15.5	11.3	22.2	20.3
Gd	2.00	4.18	6.36	6.91	3.23	4.87	3.81	1.67	2.73	2.20	2.38	1.84	6.00	6.12
Hf	5.41	2.84	3.77	4.54	1.96	3.07	2.65	0.87	1.30	1.22	1.40	2.49	4.55	4.58
Ho	0.54	0.84	1.09	1.18	0.90	1.64	1.13	0.62	0.95	0.71	0.77	0.48	1.73	1.69
La	8.72	9.56	25.36	30.30	4.36	3.49	4.64	1.59	2.40	1.86	2.05	8.54	10.47	10.86
Lu	0.38	0.23	0.29	0.32	0.36	0.69	0.43	0.28	0.42	0.32	0.33	0.20	0.64	0.66
Mo	17.20	4.24	9.29	3.85	3.16	26.10	9.07	0.89	0.83	2.15	1.94	13.00	0.72	1.99
Nb	5.78	13.50	34.50	47.60	3.83	2.53	3.04	0.89	0.97	0.84	0.99	1.82	7.37	12.30
Nd	9.60	12.34	26.28	30.87	9.02	10.56	10.53	3.67	6.76	5.40	5.45	7.75	19.14	19.76
Ni	33.7	95.8	82.2	28.7	105	105	81.9	279	174	291	238	54.2	63.3	51.6
Pb	15.5	2.68	2.33	3.44	1.21	1.17	17.40	1.07	1.36	2.02	2.02	5.39	3.03	1.13
Pr	2.60	2.65	6.16	7.39	1.75	1.91	2.07	0.71	1.28	0.99	1.00	2.04	3.91	4.01
Rb	132	7.05	23.6	21.5	1.14	0.99	1.49	8.10	8.58	8.75	1.77	2.84	12.0	1.05
Sb	0.47	0.09	0.11	0.13	0.18	0.36	0.37	0.20	0.13	0.10	0.14	0.23	0.28	0.08
Sc	2.9	17.2	15.3	13.6	27.7	27.6	20.1	18.7	23.8	21.8	20.0	7.1	22.2	19.1
Sm	2.53	3.77	6.51	7.29	2.84	4.11	3.60	1.27	2.37	1.95	1.89	1.72	5.60	5.68
Sn	2.54	1.57	2.40	2.40	1.01	1.54	1.30	0.45	0.56	0.59	0.73	0.84	1.93	1.88
Sr	9.8	98.9	182	170	48.7	83.8	339	182	187	175	71.9	383	66.3	43.5
Ta	0.57	0.98	2.51	3.54	0.31	0.20	0.24	0.07	0.08	0.07	0.08	0.24	0.55	0.88
Tb	0.37	0.71	1.00	1.06	0.61	1.00	0.71	0.34	0.55	0.43	0.46	0.33	1.05	1.10
Th	8.78	1.22	2.71	3.78	0.37	0.20	0.68	0.15	0.22	0.16	0.37	1.52	1.85	1.21
Tm	0.29	0.27	0.33	0.38	0.35	0.66	0.45	0.26	0.41	0.29	0.33	0.21	0.67	0.66
U	2.16	0.41	0.86	1.23	0.10	0.09	0.20	0.05	0.06	0.06	0.13	0.60	0.48	0.35
V	4.93	182	203	200	315	380	197	131	198	178	152	10.8	284	234

W	0.61	0.24	0.50	0.34	0.20	0.61	0.28	0.17	0.41	0.19	0.26	0.35	0.48	0.38
Y	13.2	21.7	28.7	31.4	23.9	41.6	28.6	16.4	23.2	18.7	21.6	14.0	43.3	43.9
Yb	2.35	1.54	2.03	2.35	2.27	4.26	2.79	1.66	2.74	1.92	2.12	1.33	4.55	4.52
Zn	5.1	111	128	143	88.6	115	198	57.1	69.5	61.9	62.1	3.86	106	80.7
Zr	198	106	157	197	75	115	113	36.0	51.5	47.6	55.2	68.6	185	202

sample	NOU 4	NOU 3	NOU 2	PdF 4	PdF 3	PdF 2	CONC 1	PYR	KATIR1	PNDH1
lithol	rhyolite	andésite	rhyolite	andesite	basalte	basalt	basalt	dacite	ignimbrit	andesite
Loc.	Tina	Tina	Tina	Pt des Fr	Pt des Fr	Pt des Fr	Conception	Col Pirogue	old RT1	Poindah
Terrane	Noumea	Noumea	Noumea	Noumea	Noumea	Noumea	Noumea	Noumea	Noumea	
SiO2	78.65	70.86	82.940	67.84	48.370	49.14	53.87	64.21	74.09	52.89
TiO2	0.21	0.59	0.22	0.28	1.20	0.6	0.8	0.61	0.26	1.36
A1203	10.55	13.05	8.9	13.49	15.05	14.48	14.40	16.77	12.69	15.28
Fe203	2.59	4.99	0.78	5.28	8.95	10.24	8.11	5.22	3.80	11.51
MnO	0.03	0.15	0.0	0.27	0.24	0.23	0.12	0.1	0.07	0.18
MgO	0.36	0.52	0.04	0.52	4.80	7.14	5.95	0.58	1.50	4.41
CaO	0.06	0.72	0.05	1.45	8.60	8.79	6.58	1.56	0.05	6.03
Na20	4.12	5.43	5.08	4.43	4.62	3.95	2.79	7.82	4.21	3.51
K20	1.42	1.82	0.03	3.36	1.23	0.51	3.99	0.62	1.82	0.27
P205	0.00	0.15	0.0	0.03	0.35	0.29	0.49	0.25	0.05	0.22
LOI	1.74	1.32	0.58	2.75	6.29	4.33	2.46	1.87	1.82	3.56
Total	99.74	99.59	98.62	99.69	99.70	99.70	99.56	99.61	100.35	99.22
As	0.37	0.40	0.15	1.46	0.44	0.41	1.27		5.09	1.196
Ba	581	444	49.1	803	934	483	851	202	505	78
Be	1.87	4.45	0.56	3.34	1.05	0.0	1.36		3.15	< L.D.
Bi	0.04	0.01	0.04	0.14	0.02	0.04	0.08		0.15	< L.D.
Cd	0.15	0.33	0.12	0.52	0.13	0.11	0.27		0.8	< L.D.
Ce	100.3	179.9	116.1	197.9	36.15	18.53	57.83	42.3	210	18.14
Co	2.47	3.21	0.89	0.68	25.6	35.3	23.1	9.95	0.288	26.34
Cr	39.4	49.6	53.3	39.9	40.5	347	253	46.9	5.98	107
Cs	0.31	0.22	0.23	0.61	0.96	0.76	1.17	1.01	0.20	0.24
Cu	6.65	5.31	5.8	4.48	50.6	102	85.4		3.043	123.9
Dy	7.55	12.57	8.32	12.59	4.27	2.57	3.92	0.476	18.8	4.373
Er	3.3	6.1	4.0	6.49	2.38	1.42	1.69	0.191	9.29	2.621
Eu	2.24	4.1	2.1	4.44	1.66	1.02	2.07	3.57	5.38	1.313
Ga	15.6	25.9	6.91	31.6	14.2	14.5	15.8		31.64	19.21
Gd	8.25	14.62	9.71	14.56	4.67	2.67	5.47	0.487	21.11	4.166
Hf	8.35	17.7	7.79	18.7	3.18	1.31	4.08	0.619	18.82	2.269
Ho	1.41	2.43	1.60	2.49	0.97	0.537	0.75	1.279	3.429	0.892
La	46.4	84.15	55.7	90.5	15.7	8.02	26.9	20.5	129	7.21
Lu	0.51	0.82	0.54	0.97	0.41	0.239	0.25	3.029	1.16	0.41
Mo	14.20	19.6	20.9	14.7	1.36	1.86	2.49		9.616	3.53
Nb	56.0	113	52.2	126	9.92	2.15	9.92	8.45	105.2	1.96
Nd	47.33	86.43	56.58	95.62	20.99	11.6	31.12	19.17	125.2	13.19
Ni	33.7	45.6	44.2	35.5	26.1	54.6	48.6	10.8	25.14	57.59
Pb	7.84	9.05	2.98	15.8	4.30	3.92	7.05	5.06	13.7	1.9
Pr	11.53	21.72	14.7	24.31	4.83	2.62	7.35	5.07	32.4	2.73
Rb	23.6	32.2	2.22	64.8	22.4	11.7	99.5	23.6	32.3	3.76
Sb	0.25	0.21	0.14	0.42	0.11	0.05	0.22		0.613	< L.D.
Sc	1.7	5.6	1.8	4.1	19.2	25.1	19.7			
Sm	9.97	17.34	11.76	17.96	4.95	2.94	6.56	3.912	26.03	3.78
Sn	1.62	4.25	1.9	5.91	1.18	0.77	1.25		7.973	0.90
Sr	77.6	67.4	67.4	59.1	477	803	1103	683	90	278
Ta	4.17	8.38	4.02	8.76	0.77	0.14	0.71	5.58	7.83	0.15
Tb	1.35	2.25	1.48	2.21	0.76	0.42	0.75	2.47	3.28	0.688
Th	9.12	13.8	8.71	11.1	2.42	1.06	4.15	1.43	13.43	1.09
Tm	0.51	0.84	0.55	0.91	0.36	0.22	0.27	0.19	1.29	0.386
U	2.78	4.21	2.68	3.42	0.69	0.37	1.39	1.43	4.04	0.326
V	10.4	27.0	8.07	3.06	257	242	215	144.4	< L.D.	351.2
Y	35.2	64.1	42.7	66.9	24.9	14.3	18.9	13.164	101	24.5
Yb	3.39	5.78	3.71	6.38	2.40	1.50	1.62	0.205	8.21	2.62
Zn	68.4	150	11.9	175	85.1	72.9	70.4	14	212.5	100.6
Zr	302	684	270	851	126	49	166	129.4	749.4	77.82

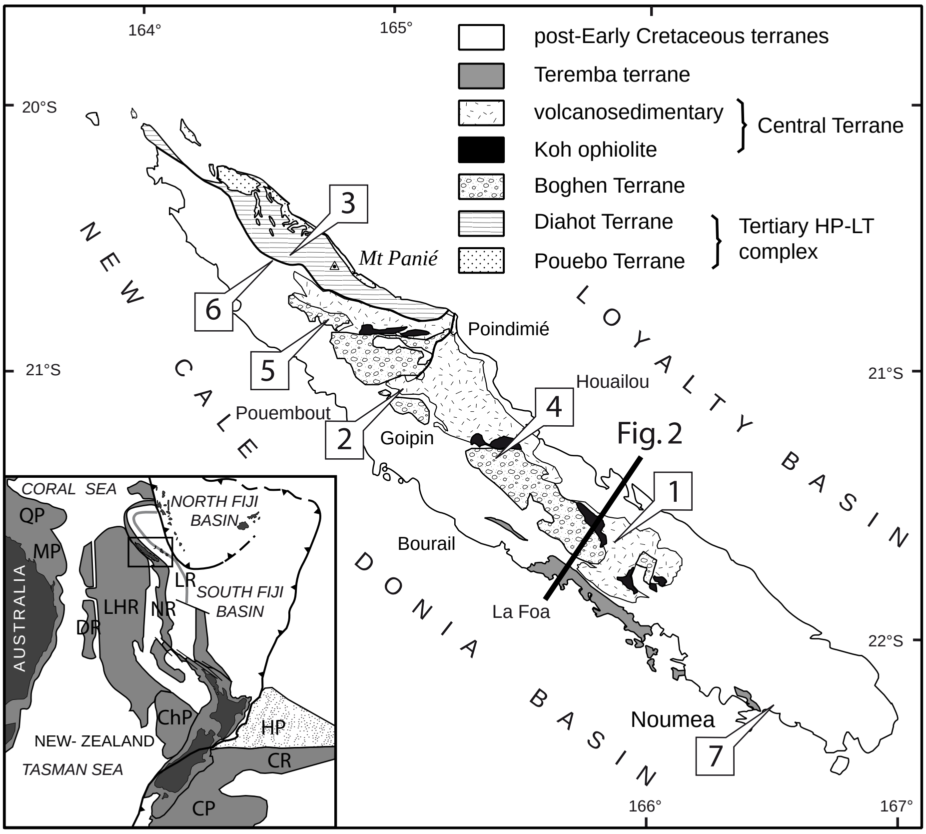

Figure 1: Geological map of New Caledonia basement terranes and U-Pb geochronology sampling localities

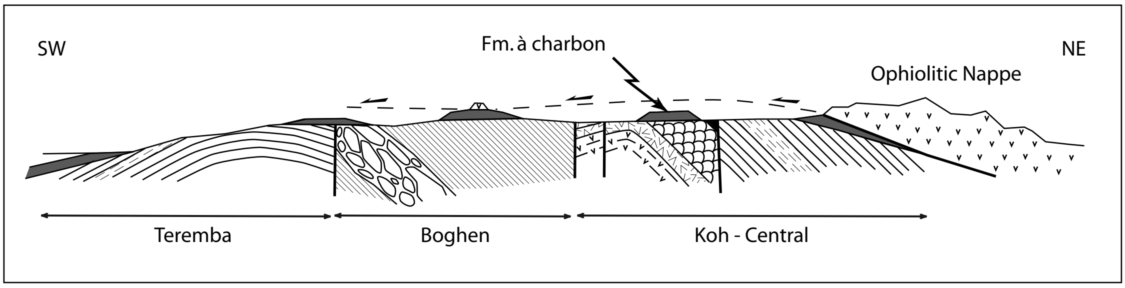

Figure 2: Very simplified cross section of New Caledonia to show the pre-Late Cretaceous terranes overlain by the unconformable "Formation à charbon"

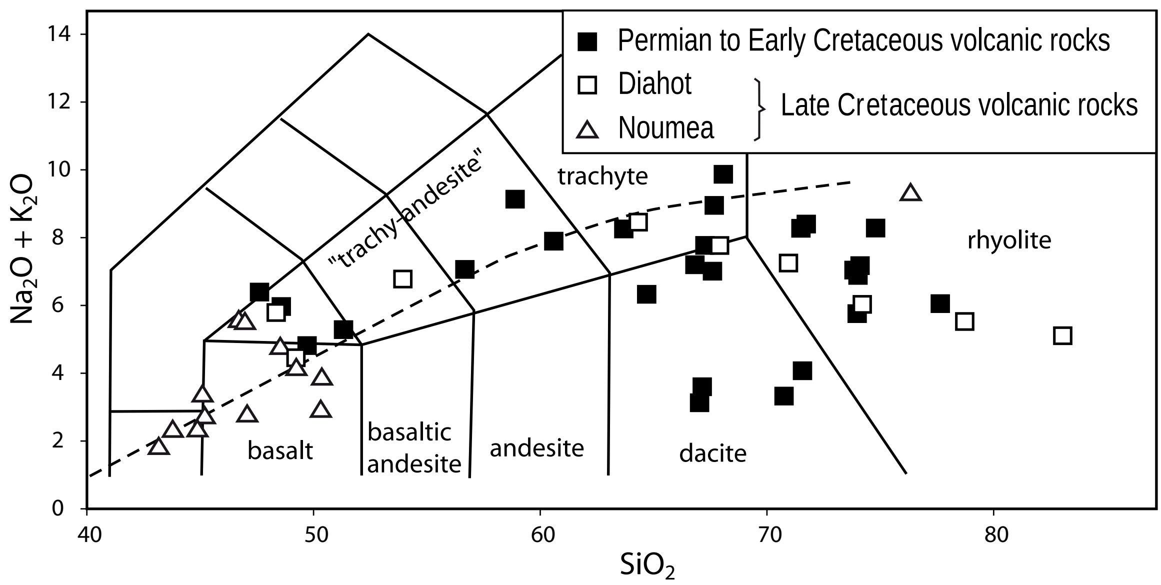

Figure 3: Total alkali vs. silica diagram

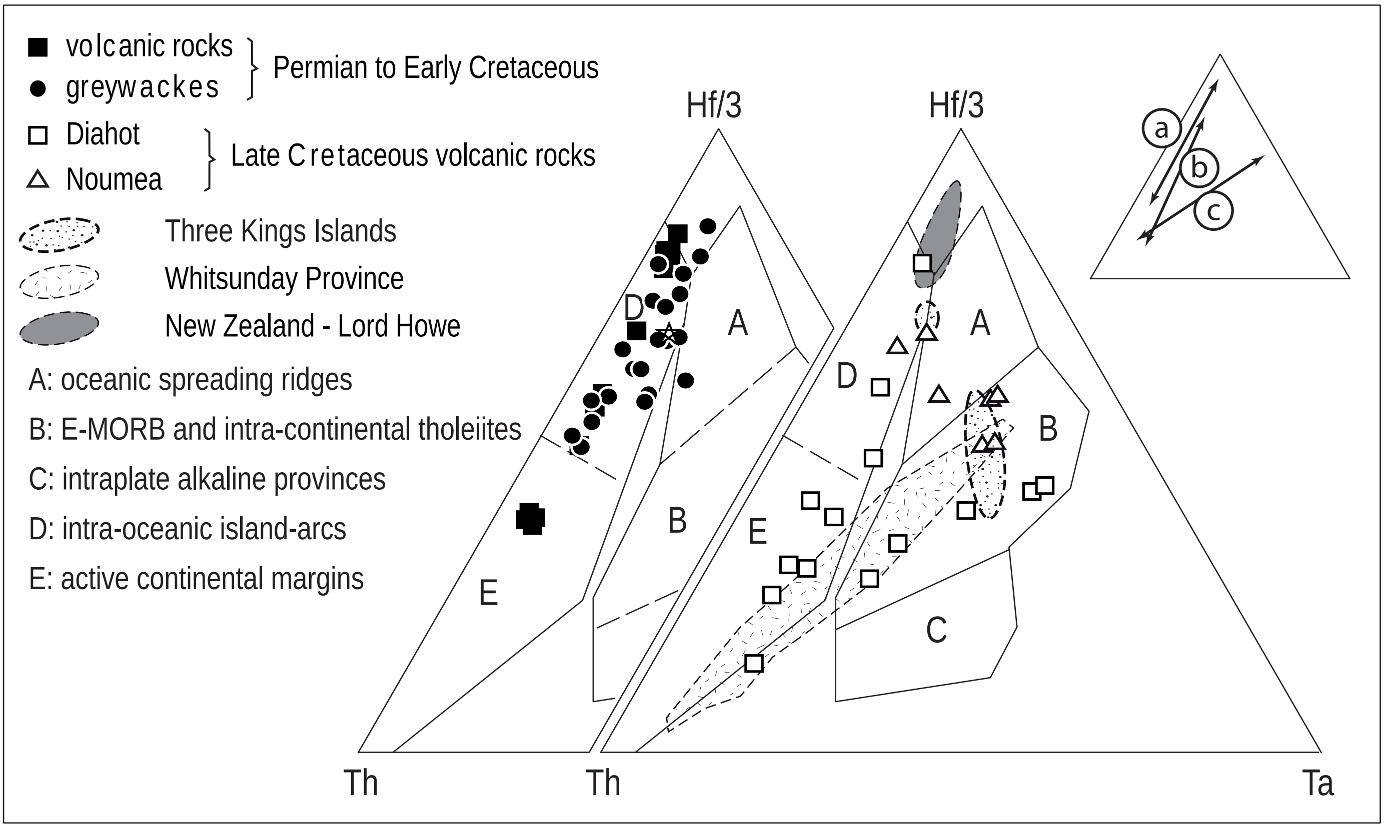

Figure 4: Hf - Th- Ta triangular diagram of Wood

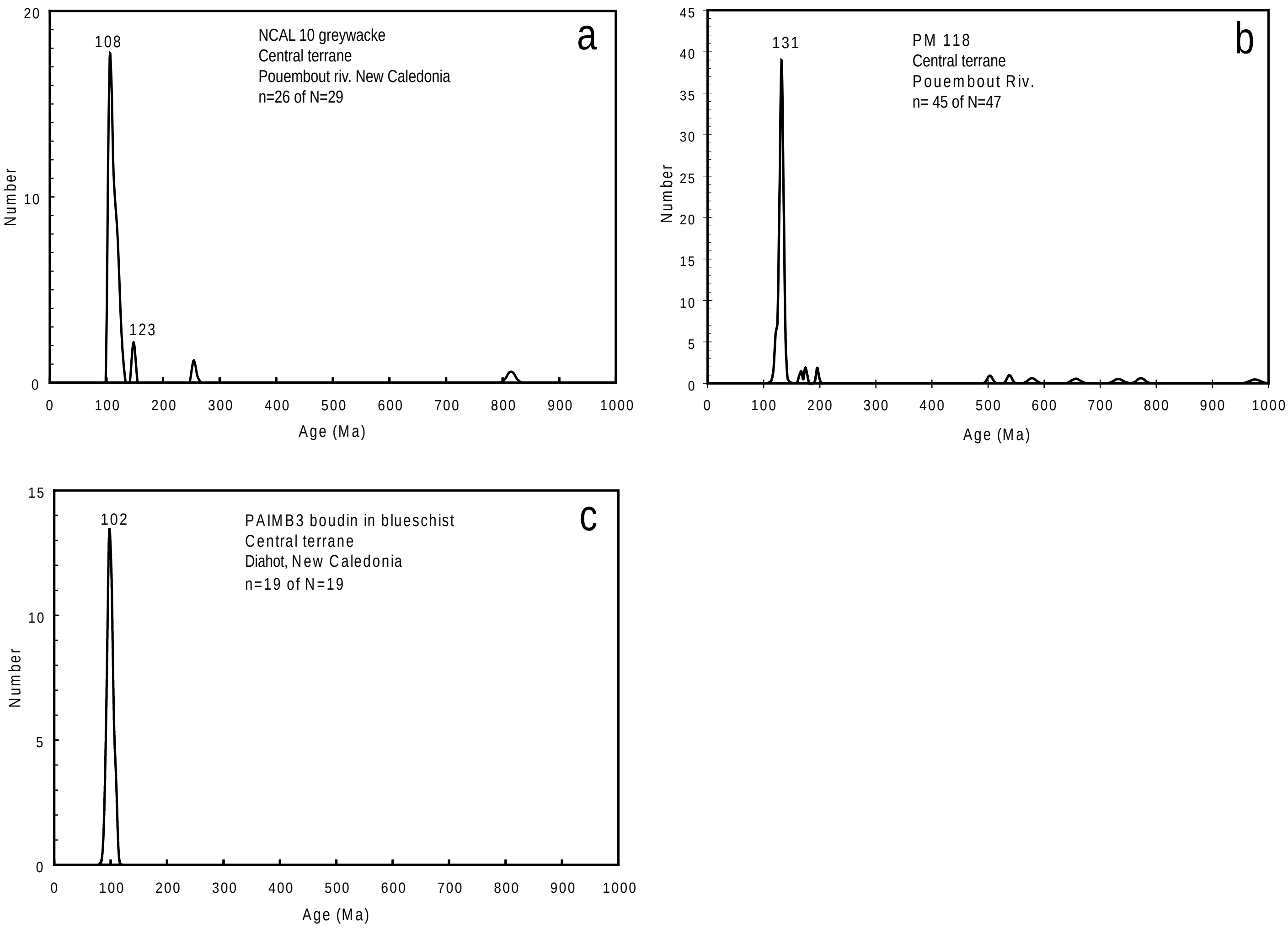

Figure 5: Probability density diagrams of detrital zircon populations from Early Cretaceous metagreywackes and greywackes from the Koh-Central terrane

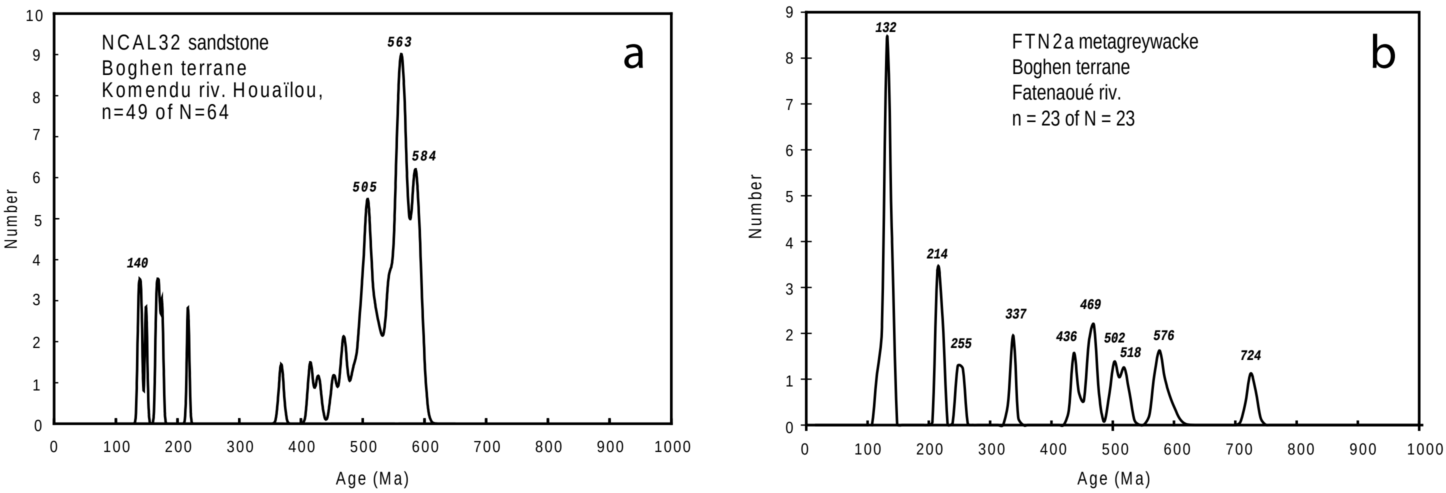

Figure 6: Probability density diagrams for detrital zircon populations from greywacke and sandstone of the Boghen terrane

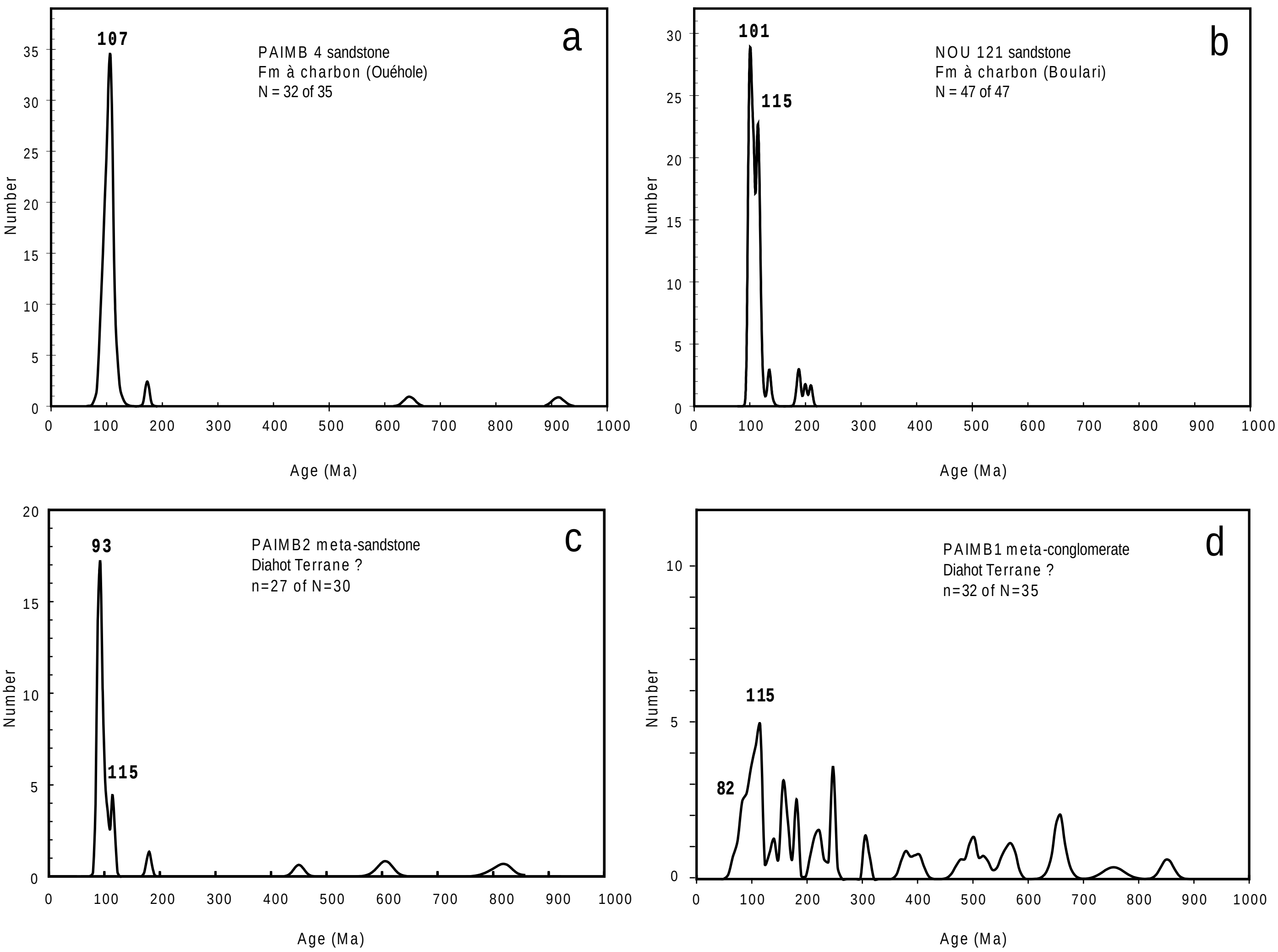

Figure 7: probability density diagram for detrital zircon populations from the Late cretaceous "Formation à charbon" and Diahot Terrane.

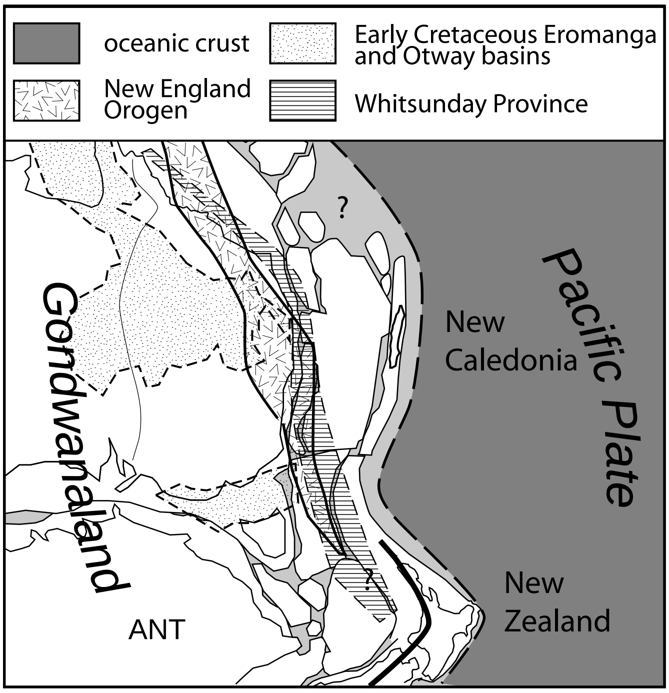

Figure 8: Reconstruction of the Gondwanaland margin in the mid-Cretaceous.

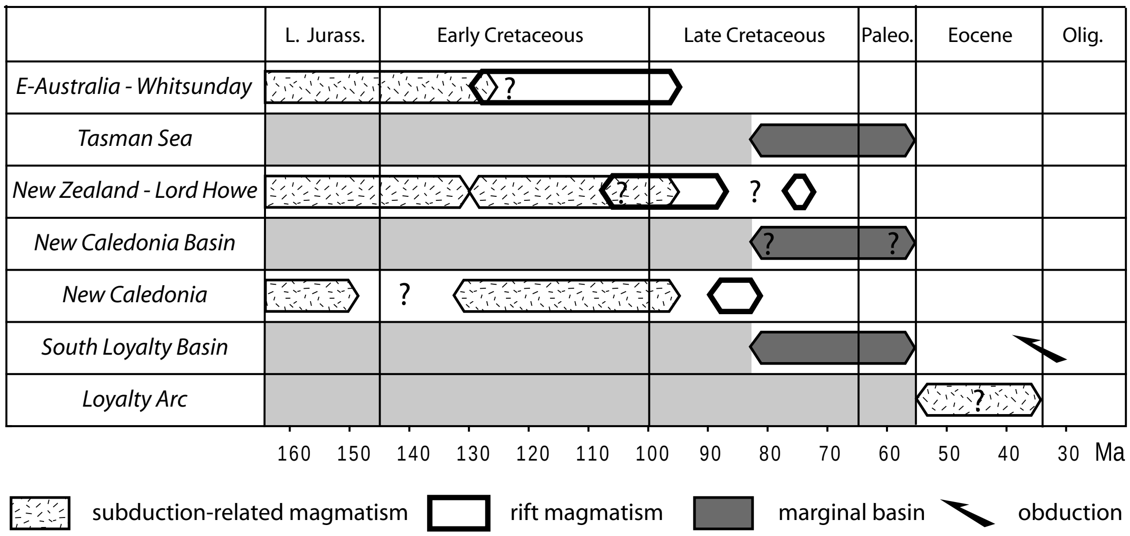

Figure 9: a summary of the chronological evolution of the Australian margin since the Jurassic

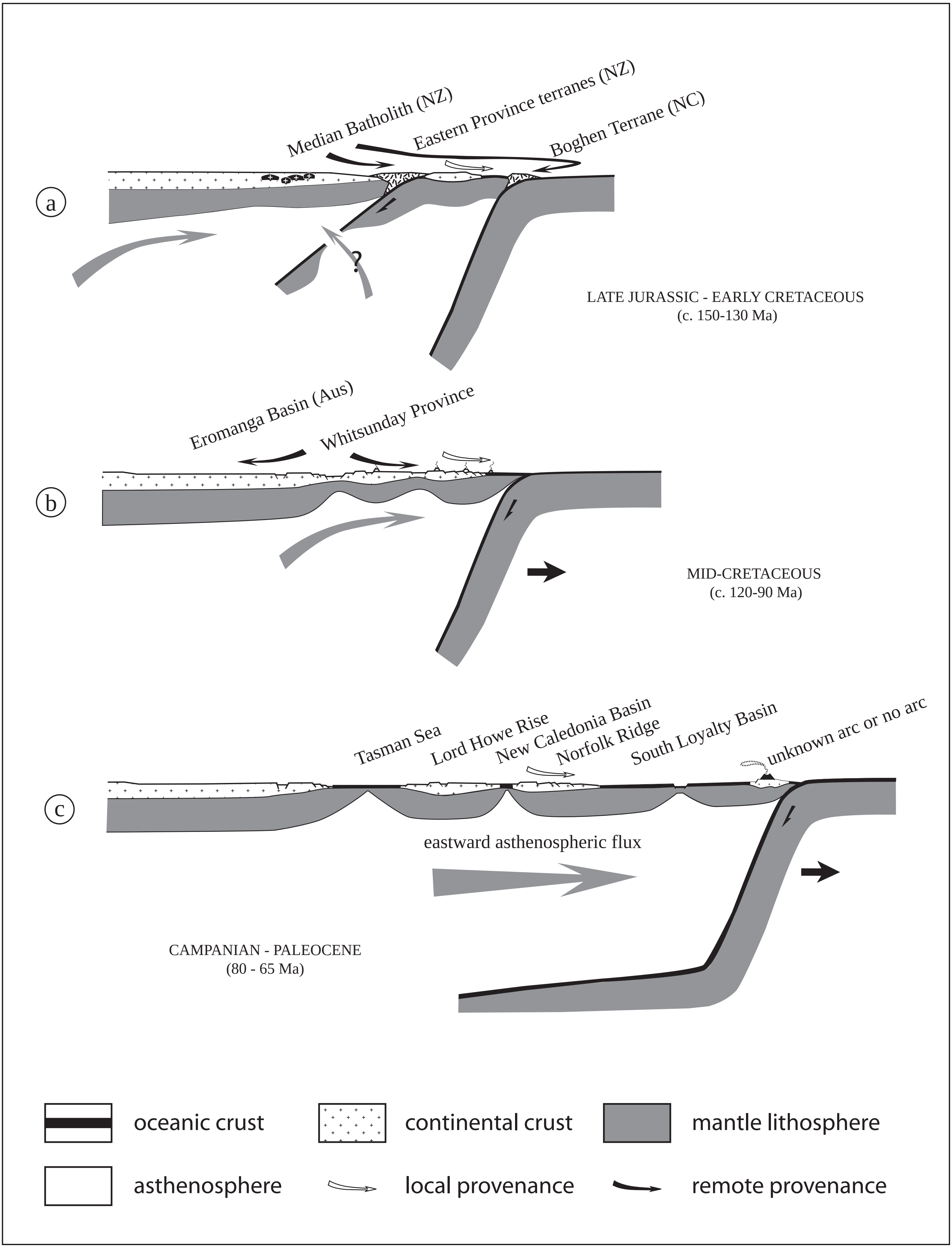

Figure 10: a tentative model for the evolution of the Australian margin